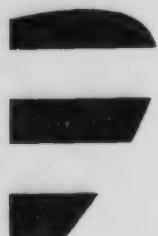


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**Soil Properties and GPR Detection of Landmines
*A Basis for Forecasting and Evaluation of GPR
Performance***

Golder Associates Ltd.

Scientific Authority:

D. Benson

Defence Research Establishment Suffield

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SOIL PROPERTIES AND GPR DETECTION OF LANDMINES
A Basis for Forecasting and Evaluation of GPR Performance



**FINAL REPORT
SUBMITTED TO:
Department of National Defense**



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REPORT ON

**SOIL PROPERTIES AND GPR DETECTION OF LANDMINES
A Basis for Forecasting and Evaluation of GPR Performance**

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SOIL PROPERTIES AND GPR DETECTION OF LAND MINES

A Basis for Forecasting and Evaluation of GPR Performance

1.0 INTRODUCTION

Recent peacekeeping missions by Canadian Forces in Somalia and the former Yugoslavia have highlighted deficiencies in current land mine detection capabilities. In response, the Improved Land Mine Detection Project (ILDMP) was initiated by the Canadian Department of National Defense (DND) in 1994 to develop and test a remote controlled, vehicle-mounted mine detection system employing multiple sensors and real-time data fusion technology. In addition to pulsed induction electromagnetic (EM), infrared (IR) and thermal neutron activation (TNA), the primary sensor suite includes a commercially developed ultra wideband ground penetrating radar (GPR). In principle, by sensing the dielectric contrast between mines and the soil matrix, GPR is capable of detecting non-metallic as well as metallic landmines. However, preliminary testing of the GPR at Defense Research Establishment Suffield (DRES) has established that GPR detection performance is strongly influenced by space and time-variable soil conditions.

As requested, Golder Associates has undertaken a combined laboratory-field investigation of DRES mine-lane soils to characterize both spatial variability in electrical properties along the test lane and time-dependent fluctuation related to environmental variables, including precipitation and temperature. In addition, an experimental test bed has been constructed to investigate the influence of climate-driven soil electrical property variation on the GPR signature of four standard land mine surrogates.

The following report presents a review of pertinent electromagnetic theory, a description of experimental procedures, resulting data and a discussion of findings and recommendations.

1.1 THE RADAR EQUATION AND SOIL PROPERTIES

It is our experience that assessment of GPR detection performance is effectively investigated within the framework of the standard radar equation

$$P_r = \frac{P_t \nu_1^2 R(f, \phi_t, \phi_r)}{(4\pi)^3 f^2 r_t^2 r_r^2} G_t(f, \theta_t) G_r(f, \theta_r) e^{-\alpha(r_t + r_r)} \quad (1)$$

as illustrated in Figure 1 (Ulaby, et al., 1982). Here, the scattering target is assumed to be a land mine, the upper half-space air and the lower half-space a uniform host soil, having effective radar propagation velocity v . Antenna gain functions $G_t(f, \theta_t)$ and $G_r(f, \theta_r)$ quantify antenna-ground coupling efficiency and the directivity or radiation pattern of the emitted (received) radar wavelet. The intrinsic attenuation rate α characterizes the influence of frequency dependent loss mechanisms that attenuate

and distort the radar wavelet during transmission through the soil. And, finally, the effective reflectivity $R(f, \theta_t, \theta_r)$, encompasses the targets frequency dependent capture cross-section or effective area, intrinsic impedance contrast and re-radiation gain function, quantifying the directivity or re-radiation pattern of the scattered radar wavelet.

To assess the detectability of a known target, we specify its electrical and geometrical characteristics. Then, with knowledge or estimation of system performance $(P_r/P_t)_{min}$, and upon prescribing certain other parameters on the basis of field measurements or reasonable assumptions, we derive corresponding limits on the remaining parameter. For instance, given representative physical properties for a given soil, we could solve for the maximum detection range. Alternatively, if we have knowledge of maximum deployment depth, we may solve the radar equation for the maximum permissible attenuation rate. On the other hand, given reasonable in-situ estimates of soil electrical properties, system performance and maximum deployment depth, we can solve the radar equation for constraints on the effective reflectivity or scattering cross-section of detectable land mines.

Furthermore, accepting that detection is a prerequisite for recognition or discrimination, let's assume that a particular land mine is detectable under a specified range of host soil conditions. Again, the radar equation is a powerful heuristic tool. Given an appropriate reflectivity function $R(f, \theta_t, \theta_r)$ for a specific land mine, the radar equation predicts the related radar signature as a function of range for specified frequency and host soil electrical properties.

However, it is not our intention here to use the radar equation to establish the signature or detection criteria for a given land mine. Rather, our goal is to illustrate the fundamental influence of soil material properties on GPR detection and discrimination of land mines.

With exception of geometrical spreading, virtually all integral mechanisms of the radar process, as described by equation (1), depend explicitly on frequency and the electromagnetic properties of the host soil: electrical conductivity σ , electrical permittivity ϵ and magnetic permeability μ . Moreover, these parameters are intrinsically frequency dependent or dispersive. Consequently, it follows from the radar equation that the signature of a given land mine can vary considerably with changing soil properties, both spatially, as related to geographically variable soil typology or localized topography-vegetation effects, and temporally, in connection with fluctuating precipitation, temperature and evapotranspiration rates.

As a result, an effective land mine detection-discrimination system must be adaptive. In particular, target detection and recognition software must be real-time adjusted to account for changing soil electromagnetic properties and resulting mine signature variation. Moreover, accepting that certain soil conditions will fall outside the range of adaptability, characterization of soil property variability as related to standard soil classification is necessary in connection with forecasting of GPR system

performance.

The following section provides a theoretical development of soil electromagnetic properties with emphasis on their practical interpretation in relation to the fundamental principles of soil science. Beyond an immediate role in connection with the assessment of laboratory and field data presented in subsequent sections, the following review provides a foundation for a more general characterization of soil electrical properties as related to standard soil classification schemes.

2.0 EM THEORY - MATERIAL PARAMETERS

We begin with the generalized Faraday's law

$$\nabla \times \mathbf{E} = -\frac{\partial \mathbf{B}}{\partial t}, \quad (2)$$

where $\mathbf{E}$ and $\mathbf{B}$ denote electrical and magnetic components of an integral vector electromagnetic field. The corresponding frequency domain relation,

$$\nabla \times \mathbf{E} = -i 2\pi f \mathbf{B}, \quad (3)$$

follows from the derivative theorem of the Fourier transform,

$$\mathbf{E}(\mathbf{r}, f) = \mathcal{F}\{\mathbf{E}(\mathbf{r}, t)\} = \int_{-\infty}^{+\infty} \mathbf{E}(\mathbf{r}, t) \exp -i 2\pi f t \, dt$$

$$\mathcal{F}\left\{\frac{\partial}{\partial t} \mathbf{E}(\mathbf{r}, t)\right\} = -i 2\pi f \mathcal{F}\{\mathbf{E}(\mathbf{r}, t)\} = -i 2\pi f \mathbf{E}(\mathbf{r}, f),$$

where $\mathcal{F}\{ \}$ denotes the Fourier transformation of an arbitrary functional argument¹. On introducing the magnetic constitutive relation

$$\mathbf{B} = \mu \mathbf{H}, \quad (4)$$

relating magnetic flux $\mathbf{B}$ to the inducing magnetic field $\mathbf{H}$, via the magnetic permeability μ , we obtain

$$\nabla \times \mathbf{E} = -i 2\pi f \mu \mathbf{H}. \quad (5)$$

Similarly, Maxwell's generalization of Ampere's law is

$$\nabla \times \mathbf{H} = \mathbf{J} = \mathbf{J}_c + \mathbf{J}_d$$

¹ Although it is customary to denote time- and frequency-domain variables by lower and upper case symbols, respectively, here we retain upper case symbols for fields and fluxes in both domains to be consistent with common representation.

$$\begin{aligned}
&= \sigma \mathbf{E} + \frac{\partial \mathbf{D}}{\partial t} \\
&= \left(\sigma + \epsilon \frac{\partial}{\partial t} \right) \mathbf{E},
\end{aligned} \tag{6}$$

where net current density $\mathbf{J}$ comprises contributions from both conduction currents, as per Ohm's Law

$$\mathbf{J}_c = \sigma \mathbf{E}, \tag{7}$$

where σ denotes electrical conductivity, and displacement currents

$$\mathbf{J}_d = \frac{\partial \mathbf{D}}{\partial t} = \epsilon \frac{\partial \mathbf{E}}{\partial t}, \tag{8}$$

arising in connection with electronic, ionic and orientational polarization in accordance with the constitutive relation

$$\mathbf{D} = \epsilon \mathbf{E}, \tag{9}$$

relating electric displacement $\mathbf{D}$ to the polarizing electric field $\mathbf{E}$, via the electrical permittivity ϵ . Applying the derivative theorem of the Fourier transform to (6), we obtain the frequency domain expression of Ampere's law

$$\nabla \times \mathbf{H} = (\sigma + i 2\pi f \epsilon) \mathbf{E} \tag{10}$$

Now, taking the curl of (5), substituting (10) and making use of the vector identity $\nabla \times (\nabla \times \mathbf{E}) = \nabla (\nabla \cdot \mathbf{E}) - \nabla^2 \mathbf{E}$, we obtain

$$\begin{aligned}
\nabla \times (\nabla \times \mathbf{E}) &= -i 2\pi f \mu (\nabla \times \mathbf{H}) \\
&= -i 2\pi f \mu (\sigma + i 2\pi f \epsilon) \mathbf{E}, \\
\nabla^2 \mathbf{E} - \nabla (\nabla \cdot \mathbf{E}) &= i 2\pi f \mu (\sigma + i 2\pi f \epsilon) \mathbf{E},
\end{aligned} \tag{11}$$

and, similarly,

$$\begin{aligned}
\nabla \times (\nabla \times \mathbf{H}) &= (\sigma + i 2\pi f \epsilon) (\nabla \times \mathbf{E}) \\
&= -i 2\pi f \mu (\sigma + i 2\pi f \epsilon) \mathbf{H}, \\
\nabla^2 \mathbf{H} - \nabla (\nabla \cdot \mathbf{H}) &= i 2\pi f \mu (\sigma + i 2\pi f \epsilon) \mathbf{H}.
\end{aligned} \tag{12}$$

Finally noting that the magnetic field is strictly non-divergent $\nabla \cdot \mathbf{H} = 0$ (no magnetic monopoles) and assuming zero net free charge distribution $\nabla \cdot \mathbf{E} = 0$, the foregoing results (11) and (12) reduce to the well known Helmholtz equations for electromagnetic waves and are commonly rewritten

$$\nabla^2 \mathbf{E} = \gamma^2 \mathbf{E} \tag{13}$$

$$\nabla^2 \mathbf{H} = \gamma^2 \mathbf{H}, \quad (14)$$

where

$$\begin{aligned} \gamma &= [i 2\pi f \mu (\sigma + i 2\pi f \epsilon)]^{1/2} \\ &= [i 2\pi f \mu \sigma + (i 2\pi f)^2 \mu \epsilon]^{1/2} \\ &= 2\pi f \mu^{1/2} \left(i \frac{\sigma}{2\pi f \epsilon} - 1 \right)^{1/2} \end{aligned} \quad (15)$$

is the so-called propagation constant. In effect, the propagation constant completely describes the propagation, attenuation and dispersion of the wavefield as controlled by intrinsic electromagnetic material properties of the medium: magnetic permeability μ , electrical conductivity σ and electrical permittivity ϵ .

2.1 ELECTRICAL CONDUCTIVITY

Electrical conductivity σ is a measure of a medium's capacity to support and sustain a free-charge conduction current density $\mathbf{J}_c = \sigma \mathbf{E}$ in the presence of an electric field $\mathbf{E}$. Conduction in soils is principally ionic, involving the net migration of electrolyte ions through interconnected pore spaces within the soil matrix as depicted in Figure 2. In comparison, electronic conduction through the solid phase soil matrix is a secondary effect. Principal soil parameters controlling ionic conductivity include porosity, moisture content and the ionic strength of the pore water.

Porosity is defined as the fractional pore volume of the soil

$$\phi = V_f/V_t = (V_a + V_w)/(V_a + V_w + V_s), \quad (16)$$

where V_a , V_w , $V_f = V_a + V_w$ and V_s denote the volume of air, water, fluids (air+water) and solids, respectively, as a fraction of total soil volume $V_t = V_f + V_s$. In practice, porosity is commonly determined via the relation $\phi = 1 - \rho_b/\rho_s$, where ρ_b denotes the bulk dry density of the soil and ρ_s is the mean matrix grain density (Hillel, 1980; Brady, 1974). Porosity is closely related to soil texture or grain size distribution and composite structural aggregation. Textural classification is based upon the relative proportions of constituent grain-size fractions (Figure 3a,b). In general, well sorted (poorly graded) fine grained soils possess higher porosity (Figure 3c,d). Additional factors include relative compaction and consolidation of the soil matrix and the presence of significant organic matter and other cementing agents, including calcium carbonate and iron oxides. Finally, in clay soils porosity can change significantly with moisture content due to relative swelling and contraction of clay aggregates (Hillel, 1980).

Partial saturation S_w is defined as the volume fraction of water relative to total pore volume or, equivalently, total fluid volume

$$S_w = V_w/V_f = V_w/(V_a + V_w). \quad (17)$$

Soil water content, or partial saturation characterizes a dynamic equilibrium condition between available water and the so-called matric potential or "suction", quantifying the soil's capacity to attract and retain moisture. As expected, soil suction varies considerably with instantaneous moisture content and is closely related to surface area and, thus, soil texture (Figure 4a). At low partial saturations, suction is due primarily to texture-dependent surface adsorption. At higher moisture levels, pore space capillarity is the dominant mechanism for water retention (Hillel, 1980; Kohnke, 1968). In general, due to vastly greater surface area per unit mass (specific surface), added clay content leads to a higher suction at any moisture level (Figure 4b). In addition, matric potential at a given moisture level depends on saturation history as illustrated in Figure 4c. The suction at a given partial saturation is lower during wetting (sorption) than during drainage (desorption). In general, this hysteresis effect is related to the relative geometrical distribution of fluid (air/water) within the pore volume and is more pronounced for coarse textured soils at high partial saturations.

Soil water is classified according to its relative retention as gauged by matric potential. Adsorbed or hygroscopic water is surface bound water at matric potential exceeding 31 bar, constituting a thick film (several molecular layers) coating individual grain surfaces. In contrast, capillary water is retained via surface tension at matric potentials between 0.1 and 31 bar and extend substantially into the pore volume, depending on grain contact geometry as illustrated in Figure 4b. Finally, water residing within the pore volume at a matric potential lower than 0.1 bar is effectively free to drain under the influence of gravity and, consequently is referred to as free or gravitational water. In practice, water available to support vegetation is restricted by drainage and retention. Free water drains from the soil too readily and adsorbed water is bound too tightly to be extracted. Consequently, capillary water constitutes the bulk of water available to support vegetation. Specifically, as illustrated in Figure 5, capillary water between matric potentials of 0.1 bar (field capacity) and 15 bar (wilting point) is defined as available water. Note that although clays possess both the highest field capacity and wilting point, silt loams accommodate more available water.

Finally, pore water electrical conductivity depends on the combined concentration of dissolved salts and colloidal solids. The principal ionic constituents of natural ground water are tabulated in Figure 6a with a selection of corresponding ionic mobilities at 25° C in Figure 6b. Bulk ionic conductivity is

$$\sigma_w = 9.65 \times 10^4 \sum_{i=1}^n n_i \mathcal{M}_i, \quad (18)$$

where n_i and $\mathcal{M}_i$ denote the valence adjusted concentration and ionic mobility, respectively, of the i^{th} ionic species and $\sum_{i=1}^n$ denote the summation over the total number of species n (McNeil, 1980). Due to enhanced ionic mobility and decreasing fluid viscosity, electrical conductivity increases significantly with increasing temperature

according to the relation $\sigma_w = \sigma_{w,25^\circ C} [1 + \lambda(T - 25)]$, where temperature coefficient λ depends on the ionic composition of the soil water (McNeil, 1980). For a sodium chloride solution, $\lambda \approx 0.002$, implying a 2 percent conductivity increase per degree centigrade. A range of representative groundwater data are presented in Figures 6c,d.

Now, returning our attention to the bulk electrical conductivity of a soil, a useful empirical relation, accounting for foregoing factors, is a generalized form of Archie's Law (Keller and Frischknecht, 1966; McNeil, 1980; Keller, 1987)

$$\sigma = a\sigma_w S_w^m \phi^m, \quad (19)$$

where σ_w , S_w and ϕ denote pore water conductivity, partial saturation and porosity, respectively. Empirical constants a (1.0-1.3), m (1.3-2.0) and n (1.0-2.25) are related to matrix texture, consolidation and moisture history. Figure 7a depicts Archie's relation as a function of partial saturation for $\phi = 0.5$ and a range of parameter settings spanning typical soil values $a = 1.0$, $n=2.0$ and $m=1.4$. In essence, Archie's law stipulates that bulk soil electrical conductivity depends on total soil moisture content $\psi_w = S_w \phi$ to a degree dictated by pore water conductivity σ_w .

The porosity exponent m , is related to the dominant shape of matrix particles. Figure 7b summarizes the findings of a study by Jackson et al. (1978), defining a linear relation $m = 2.5 - 1.3\varphi$ between m and the sphericity of matrix grains φ . In practice the porosity exponent ranges from approximately $m=1.3$ for well rounded quartz sand grains to approximately $m=2.0$ for platy silt and clay particles. Similarly, the saturation exponent n depends predominantly on the distribution of soil water within the pore structure and, consequently, on the texture-controlled matric potential. As per previously described hysteresis of the matric potential, the appropriate saturation exponent can be significantly influenced by saturation history.

Further generalization of Archie's Law

$$\sigma = a\sigma_w S_w^m \phi^m + \sigma_s, \quad (20)$$

and a related formulation due to Rhoades et al. (1976)

$$\sigma = a_r \sigma_w S_w^2 \phi^2 - b_r \sigma_w S_w \phi + \sigma_s,$$

with a_r and b_r denoting texture-dependent empirical constants, recognize the additive contribution of colloidal surface conductivity σ_s , associated with anomalous ion concentrations arising in near proximity to negatively charged clay and organic colloidal surfaces.

As a result of isomorphous cation substitution within the layered silicate structure and incomplete charge neutralization at lattice terminations, clay particles possess a net negative surface charge. Similarly, amorphous organic colloids possess a pH-dependent negative charge due to dissociation of H^+ from organic functional groups

(Bohn, *et al.*, 1985). Although the resulting surface charge density is generally greater than that of clays, the result is the same. Unbalance negative charge is compensated externally via adsorption of aqueous cations within the surrounding pore water, leading to a characteristic diffuse charge distribution known as an electrostatic double layer.

As depicted in Figure 7c, the double layer of adsorbed ions comprises a fixed layer (Stern layer) immediately adjacent to the colloid surface and a diffuse layer with ionic concentration gradually diminishing toward equilibrium with the bulk electrolyte at a characteristic distance from the colloid surface

$$\Delta = \frac{1}{eZ} \left(\frac{\kappa kT}{8\pi n} \right)^{1/2} \quad (21)$$

where e denotes the elementary charge on an electron, n is the effective pore water ion concentration and Z is the related ion valence. κ represents soil water dielectric constant, k is Boltzmann's constant and T denotes temperature in Kelvin degrees (Overbeek, 1952; Hillel, 1980).

By increasing the net free charge density, the so-called diffuse double layer leads to preferential ionic conduction immediately adjacent to the colloidal surface that is accounted for via addition of the effective surface conductivity term σ_s . Figure 7d depicts the influence of surfacial conductivity as a function of partial saturation for $\sigma_s/\sigma_w = 0.1, 0.2$ and 0.3 .

As illustrated by Figure 8a, colloidal surface conductivity is the dominant charge transport mechanism for the case of low conductivity pore fluids or, for that matter, low partial saturations. However, the extent and influence of the diffuse layer on bulk electrical conductivity diminishes abruptly as pore water conductivity reaches a critical level σ_w^* dependent on soil matrix characteristics. For higher pore water conductivities, the corresponding soil conductivity is predicted by (19). We note that this transition is consistent with (21) which predicts an effective contraction of the diffuse layer with increasing ionic concentration of the soil water.

Figure 8b displays empirical data reported by Schön (1996) for a clay (a) and three sands (b,c,d). As expected, both surface conductivity σ_s and the transition pore water conductivity σ_w^* are significantly higher for the clay than for sands. However, in no case is the transition pore water conductivity lower than 1.0 mS/m, suggesting that surface conductivity remains significant over the typical range of soil water conductivity 1 - 10 mS/m. Finally, regression analysis for coarser grained fractions indicated the relation $\sigma_s = 9.0 \times 10^{-4} d^{-0.4}$ between surface conductivity and mean grain diameter d . Figure 8c presents a similar comparison of resistivity $\rho = 1/\sigma$ as a function of clay content (Keller, 1987).

The potential of a particular soil constituent as a repository of available charge carriers is quantified by the corresponding cation exchange capacity (CEC) as tabulated for selected clays and soil textural classes in Figures 9a-b, respectively. Related

specific surface data summarized in Figures 9c-e establish a direct correlation between CEC and surface area as per Figures 10a-b. Clearly, the effect of anomalous surface charge densities associated with the clay fraction of a soil is greatly amplified by its enormous relative surface area. Recall the related influence of clay content and, consequently, surface area on the moisture retention properties of soils. Moreover, as previously indicated, organic colloids can have a still greater effect. Figure 10c illustrates the particularly high CEC associated with organic colloids and the dependence on soil pH.

On application of a static electric field, charge carriers comprising the diffuse double layer, although electrostatically bound to the grain surface, are displaced parallel and anti-parallel to the applied field, yielding a net polarization of the ionic atmosphere surrounding an isolated grain as depicted in Figure 11a (Schwartz, 1962; Dukhin, 1971). Offsetting contributions of adsorbed and diffuse layers is discussed by Lyklema, *et al.* (1983). Subsequently, on removal of the field, the polarized double layer "relaxes" back to its equilibrium configuration at a rate depending principally upon effective diffusivity and grain size-geometry. In particular, the relaxation time τ is directly proportional to the square of grain radius and inversely related to the effective diffusion coefficient.

In the presence of a time-varying field, alternating polarization and relaxation of the electrostatic double layer leads to significant dispersion of the electrical conductivity at frequencies near the associated relaxation frequency $f = 1/\tau$. In general, as the rate of field fluctuation approaches and exceeds the effective relaxation frequency, a marked increase in electrical conductivity is observed. In practice, due to the broad range of grain sizes and geometries comprising a natural soil, a continuous spectrum of relaxation frequencies yields broad-band dispersion, extending well into the microwave region.

In addition to direct polarization of the double layer, interaction of colloidal surface charge accumulations can effectively constrict and reduce pore space connectivity, producing an ion selective membrane effect and associated charge polarization as depicted in Figure 11b. In particular, membrane polarization results in significant charge concentration gradients and associated diffusion potentials that oppose the polarizing field potential and effectively restrict the rate of charge transport. As charge polarization increases, so too does the opposing diffusion potential as the associated effective impedance asymptotically approaches a limiting value (Ward, 1990; Ward and Fraser, 1967). The same osmotic effects control the rate of charge relaxation between reverse polarizations and, thus, determine corresponding frequency dispersion of the electrical conductivity. The magnitude of membrane polarization effects depends on the ratio of double layer thickness Δ to effective pore radius. As expected, an increase in pore water ion concentration leads to a substantial reduction in polarizability due to related contraction of the diffuse layer.

A similar, polarization occurs when pore space connectivity is blocked by a matrix

grain as illustrated in Figure 11c. Assuming a reactionless, non-conducting mineral grain, application of a time varying electric field results in Maxwell-Wagner polarization due to accumulation of oppositely charged electrolyte ions at opposing grain interfaces. Again, relaxation of the polarized charge distribution is diffusion limited (Olhoeft, 1985; Wobschall, 1977; Hanai, 1960). More commonly, the mineral grain is reactive and possesses finite electrical conductivity so that Maxwell-Wagner polarization is accompanied by electronic conduction across the polarized grain via electrochemical reactions at grain surfaces (Ward, 1990; Ward and Fraser, 1967). Principal electrochemical polarization processes include oxidation-reduction, ion exchange and clay-organic reactions (Olhoeft, 1985; 1987). The associated rate of charge transport is ultimately limited either by the chemical reaction rate or the effective diffusivity associated with aqueous charge transport to and from the reactive interface. In general, relaxation frequencies associated with diffusion limited processes are inversely related to grain size, with colloidal size particles producing significant dispersion at frequencies exceeding 1 GHz (Figure 11d). In contrast, the dispersion characteristics associated with reaction rate or kinetics limited processes are independent of particle or pore size distributions. In practice, a range of electrochemical polarization processes are active for a given soil and ordinarily overshadow the underlying Maxwell-Wagner polarization at frequencies below 1 MHz. Conversely, at frequencies exceeding 100 MHz, pure interfacial polarization effects tend to dominate (Olhoeft, 1985; Parkhomenko, 1967).

In general, it is established on the basis of empirical data that the combined dispersion effects due to foregoing polarization processes are adequately modelled via the simple circuit analogy depicted in Figure 12a, consisting of a free electrolytic pore passage (pure resistance) in parallel with an obstructed-polarized pore passage (series resistance-reactance). In general, related dispersion of the electrical conductivity is quantitatively described via a relaxation model, incorporating a continuous time constant distribution (Hobbs, 1974; Olhoeft, 1985; Kamenetsky and Tomofeyev, 1992)

$$\sigma(f) = \sigma_{dc} \left[1 + \left(\frac{\eta}{1 - \eta} \right) \int_0^\infty \frac{i2\pi f \tau T(\tau)}{1 + i2\pi f \tau} d\tau \right]. \quad (22)$$

Here, $\eta = (\sigma_\infty - \sigma_{dc})/\sigma_\infty$ denotes the free-charge polarizability, or so-called charge-ability, where $\sigma_{dc} = \sigma'(f \ll 1/2\pi\tau)$ and $\sigma_\infty = \sigma'(f \gg 1/2\pi\tau)$ represent static and high frequency conductivity, and $T_\sigma(\tau)$ represents a continuous time constant distribution characterizing the relaxation rates of constituent free charge polarization mechanisms. Accordingly, the electrical conductivity incorporates both in-phase and quadrature components

$$\sigma = \sigma' + i\sigma''. \quad (23)$$

In particular, for a single polarization mechanism, having effective time constant $T_\sigma(\tau) = \delta(\tau - \tau_\sigma)$, where $\delta(\tau)$ denotes the Dirac delta function, the complex electrical

conductivity is given by

$$\sigma'(f) = \sigma_{dc} \left[1 + \left(\frac{\eta}{1-\eta} \right) \frac{(2\pi f \tau_\sigma)^2}{1 + (2\pi f \tau_\sigma)^2} \right] \quad (24)$$

and

$$\sigma''(f) = \sigma_{dc} \left[\left(\frac{\eta}{1-\eta} \right) \frac{2\pi f \tau_\sigma}{1 + (2\pi f \tau_\sigma)^2} \right] \quad (25)$$

Predicted dispersion is qualitatively depicted in Figure 12b. Note that quadrature conductivity $\sigma''(f)$ possesses a peak value at the relaxation frequency $f_\sigma = 1/2\pi\tau_\sigma$, with amplitude depending on the chargeability η . Figure 13 illustrates the relationship between chargeability and soil texture. Consistent with the surface related nature of free-charge polarization processes, an inverse correlation is observed between chargeability and grain size. In particular, membrane effects are closely related to clay content and distribution. However, there is evidently an optimum clay content, above which additional clay reduces chargeability. According to Ogilvy and Kuzmina (1972), maximum chargeability is observed for clay contents ranging from 3 to 8 percent by weight, depending on clay mineralogy. In effect, additional clay disturbs the balance between free and polarized pore passages as per the phenomenological model. Consequently, chargeability and related dispersion of the electrical conductivity is greatest for loamy soils. Polarizability also increases with moisture content to maximum effect at approximately 5-20 percent water by weight. Higher moisture contents lead to reduced chargeability (Iliceto, et al., 1982; Ogilvy and Kuzmina, 1972).

In general, as a consequence of the causality constraint on physically realizable systems, any dispersion of $\sigma'(f)$ must be accompanied by related dispersion of $\sigma''(f)$ as per the Kramers-Kronig relations (Kittel, 1958; Fuller and Ward, 1970)

$$\sigma''(f) = \frac{1}{\pi} \int_{-\infty}^{\infty} \frac{\sigma'(\varphi)}{\varphi - f} d\varphi = -\frac{2f}{\pi} \int_0^{\infty} \frac{\sigma'(\varphi)}{f^2 - \varphi^2} d\varphi \quad (26)$$

$$\sigma'(f) = -\frac{1}{\pi} \int_{-\infty}^{\infty} \frac{\sigma''(\varphi)}{\varphi - f} d\varphi = \frac{2}{\pi} \int_0^{\infty} \frac{\varphi \sigma''(\varphi)}{f^2 - \varphi^2} d\varphi, \quad (27)$$

providing, in theory, a means of deriving one parameter from the other ². However, as Fuller and Ward (1970) observed, practical implementation is restricted by necessity

² Mathematically, Kramers-Kronig relations define a Hilbert transform pair for real and imaginary parts of a complex transfer function. In effect, the Hilbert transform is a quadrature phase shift filter that preserves the amplitude of spectral components in the transform domain while affecting a 90 degree phase shift, positive or negative, according to sign. Singularities in the integrands are accounted for analytically by taking Cauchy principal values. In practice, computation is readily performed via the convolution operator $(-i\pi f)^{-1}$ or in the Fourier transform domain via multiplication with the complex function $i \operatorname{sgn}(t)$ (Bracewell, 1986).

to measure the real component over all frequencies. Moreover, as we shall discover, measured electrical properties are effective parameters, incorporating non-separable free-charge and bound-charge contributions.

Finally, at very high frequencies, effective frictional forces due to collisions between free ionic charge carriers ultimately prevent the oscillatory motion of electrolyte ions in response to polarizing electric field variations. In effect, associated electrical conductivity vanishes. The result is demonstrated as follows. Assuming that the instantaneous frictional force acting on a given ion is directly proportional to the mean nonrandom velocity of the ion, an approximate expression for the electrical conductivity is (King and Smith, 1981)

$$\sigma'(f) = \frac{e^2 Z_-^2 n_- \nu_-}{m_- (\nu_-^2 + 4\pi^2 f^2)} + \frac{e^2 Z_+^2 n_+ \nu_+}{m_+ (\nu_+^2 + 4\pi^2 f^2)} \quad (28)$$

$$\sigma''(f) = \frac{e^2 Z_-^2 n_- (2\pi f)}{m_- (\nu_-^2 + 4\pi^2 f^2)} + \frac{e^2 Z_+^2 n_+ (2\pi f)}{m_+ (\nu_+^2 + 4\pi^2 f^2)}, \quad (29)$$

where n_- , m_- , Z_- , ν_- and n_+ , m_+ , Z_+ , ν_+ denote the unit ion concentration, mass, valence and collision frequency for anions and cations, respectively, and e is the elementary charge. Now, consider the particular case of an electrolyte comprising an equal numbers of positive and negative ions having equivalent mass ($m_- = m_+$), valence ($Z_- = Z_+$) and collision frequency. The resulting complex electrical conductivity at a given temperature is

$$\sigma'(f) = \sigma_\infty \left[\frac{1}{1 + (2\pi f \tau_\infty)^2} \right] \quad (30)$$

$$\sigma''(f) = \sigma_\infty \left[\frac{2\pi f \tau_\infty}{1 + (2\pi f \tau_\infty)^2} \right] \quad (31)$$

where we let $\sigma_\infty = 2e^2 Z^2 n m^{-1} \tau_\infty$ denotes the previously defined high frequency conductivity and τ_∞ is the characteristic mean free time between collisions or, in effect, the time required for mean non-random velocity to become negligible following removal of the applied field. As for conductivity dispersion associated with membrane and electrochemical polarization, the foregoing relations describe a relaxation process. In contrast, however, corresponding high-frequency dispersion involves a rapid reduction in $\sigma'(f)$ and coinciding peak in $\sigma''(f)$ at the characteristic frequency $\nu_\infty = 1/\tau_\infty$. In practice, however, the characteristic frequency for typical electrolytes falls in the infrared region (Purcell, 1963)³ of the electromagnetic spectrum, indicating that

³ Representative mean free time $\tau_\infty = \sigma m / n e^2$ is estimated on the basis of the measured electrical conductivity of a NaCl solution of known concentration. For $n=0.001 \text{ g/cm}^3$ and $\bar{\sigma}=200 \text{ mS/m}$, $\tau_\infty = 1.8 \times 10^{-14} \text{ s}$, implying a relaxation frequency of $\nu_\infty = 5.5 \times 10^{13} \text{ Hz}$.

the associated loss of conductivity has little practical influence on electromagnetic propagation at radar frequencies.

In summary, the electrical conductivity of soils depends principally on soil water as determined by porosity, partial saturation and pore water conductivity. Another critical factor is the cation exchange potential of matrix constituents, particularly clays and organic colloids. In addition, free-charge polarization processes produce significant frequency dispersion of the electrical conductivity, leading to a finite phase lag between polarizing electric field variations and resulting conduction currents. In general, below 100 MHz, the in-phase component of the corresponding complex conductivity increases incrementally with frequency due to a range of free-charge polarization phenomena. The strength of membrane and higher frequency colloid-electrochemical effects depends primarily on clay content.

Figure 14a provides an indication of the general relation between soil texture and electrical conductivity. In general, soil conductivity is observed to increase as dominant grain size decreases. As predicted, the electrical conductivity measured at 300 kHz is consistently higher than lower frequency values. Finally, and principally via its effect on pore water electrical conductivity, temperature exerts a direct influence on bulk soil conductivity. As illustrated in Figure 14b, soil conductivity is observed to decrease gradually with declining temperature to 0° C, before diminishing rapidly as pore water changes state. The relatively gradual conductivity reduction for one of the clay samples may be attributable to a relatively high proportion of adsorbed water and an associated predominance of surface conductivity.

Representative empirical data on the electrical conductivity of soils are tabulated in Appendix I.

2.2 ELECTRICAL PERMITTIVITY

Electrical permittivity quantifies a medium's capacity to support and sustain a polarization or displacement current density $J_d = \epsilon \partial E / \partial t$, proportional to the rate of change in the polarizing field $\partial E / \partial t$. Thus, in contrast with conduction currents, displacement currents are generally speaking quadrature-phase with the polarizing field E . Like electrical conductivity, however, the electrical permittivity is characterized by significant dispersion and, consequently, is a complex parameter

$$\epsilon(f) = \epsilon'(f) - i\epsilon''(f), \quad (32)$$

comprising real $\epsilon'(f)$ and imaginary $\epsilon''(f)$ components, respectively quadrature-phase and in-phase with the time-varying electric field.

Compared with mechanisms for free-charge conductivity dispersion, polarization processes responsible for bound-charge permittivity dispersion are relatively well understood and documented (von Hippel, 1954, 1984; Kittel, 1967; King and Smith, 1981). The frequency dependent electrical permittivity spectrum and associated polarization mechanisms are illustrated in Figure 15. Three distinct processes contribute

to the net polarizability of a bound charge distribution: electronic, ionic and orientational (dipolar). Resulting polarization is additive as indicated in Figure 15b. Each mechanism is associated with a plateau in the polarizability spectrum, terminated by a localized dispersive zone in proximity of the corresponding resonance or relaxation frequency, above which related polarization ceases (Figure 15c).

In particular, electronic polarizability arises due to the net displacement of electrons relative to the nucleus of individual atoms, with associated resonance occurring at optical frequencies, depending on specific atomic characteristics. Ionic or atomic polarization involves non-uniform intramolecular distribution of charge due to the differing electronegativities of constituent atoms. As a result, bound atoms acquire effective charges of opposite polarity, leading to a permanent dipole moment and relative displacement or polarization upon application of an external electric field. Resonance dispersion associated with field induced polarization of the asymmetric charge distribution occurs at thermal frequencies as found previously for relaxation of free ions in solution. Finally, time dependent alignment of permanent molecular dipoles with the polarizing field gives rise to a separate orientational (dipolar) polarization, having relaxation frequencies in the microwave range and, consequently, of particular consequence for high-frequency ground radar applications.

As previously discussed in connection with free-charge polarization and associated dispersion of the electrical conductivity, bound-charge polarizability implies that the associated electrical permittivity is dispersive and, consequently, complex valued as dictated by corresponding Kramers-Kronig relations

$$\epsilon''(f) = \frac{1}{\pi} \int_{-\infty}^{\infty} \frac{\epsilon'(\varphi)}{\varphi - f} d\varphi = -\frac{2f}{\pi} \int_0^{\infty} \frac{\epsilon'(\varphi)}{f^2 - \varphi^2} d\varphi \quad (33)$$

$$\epsilon'(f) = -\frac{1}{\pi} \int_{-\infty}^{\infty} \frac{\epsilon''(\varphi)}{\varphi - f} d\varphi = \frac{2}{\pi} \int_0^{\infty} \frac{\varphi \epsilon''(\varphi)}{f^2 - \varphi^2} d\varphi. \quad (34)$$

It follows that any change in real permittivity $\epsilon'(f)$ is accompanied by related variation of the imaginary component $\epsilon''(f)$. In particular, as illustrated in Figure 15c, localized dispersion of $\epsilon'(f)$ associated with electronic, ionic and orientational polarization processes is accompanied by corresponding peaks in $\epsilon''(f)$. Over remaining portions of the permittivity spectrum, where $\epsilon'(f)$ is effectively constant, $\epsilon''(f)$ is negligible.

In practice, because the electrical permittivity or dielectric constant ⁴ of water ($\kappa \approx 80$) is more than an order of magnitude greater than the average for dry soil matrix constituents ($\kappa \approx 3-6$), soil moisture content is the primary factor controlling bulk soil electrical permittivity. Moreover, because permittivity dispersion associated with orientational polarization of the dipolar water molecule occurs at GPR frequencies, characterization of the electrical permittivity of water is essential.

⁴ Here we define the dielectric constant as $\kappa = \epsilon'(0)/\epsilon_0$, where $\epsilon'(0)$ denotes the static real electrical permittivity and $\epsilon_0 = 8.85 \times 10^{-12}$ denotes the electrical permittivity of free space.

An accepted theoretical model describing the orientational polarization of polar molecules was formulated by Debye (1929). Assumption that time-dependent orientational polarization approaches the static value $\epsilon_{dc} = \epsilon(0)$ at a rate proportional to the difference between instantaneous and static values, leads to the well known Debye relaxation spectrum (Hobbs, 1974; von Hippel, 1984; King and Smith, 1981)

$$\epsilon(f) = \epsilon_{\infty} + \frac{\epsilon_{dc} - \epsilon_{\infty}}{1 + i2\pi f\tau_e}, \quad (35)$$

or

$$\epsilon'(f) = \epsilon_{\infty} + \frac{\epsilon_{dc} - \epsilon_{\infty}}{1 + (2\pi f\tau_e)^2} \quad (36)$$

$$\epsilon''(f) = \frac{2\pi f\tau_e (\epsilon_{dc} - \epsilon_{\infty})}{1 + (2\pi f\tau_e)^2} \quad (37)$$

where τ_e denotes the characteristic relaxation time⁵ and $\epsilon_{\infty} = \epsilon'(f \gg 1/2\pi\tau_e)$ denotes the high-frequency real permittivity, including electronic and ionic contributions but excluding orientational polarization effects⁶.

The general nature of the Debye dispersion spectrum is illustrated in Figure 16a as a function of normalized frequency. Real permittivity $\epsilon'(f)$, decreases from approximately the static value ϵ_{dc} (for $2\pi f\tau_e \leq 0.1$) to the corresponding high-frequency value ϵ_{∞} (for $2\pi f\tau_e \geq 10$) over a relatively narrow frequency band surrounding the relaxation frequency. In particular, for water at 25° C, dispersion occurs largely over the frequency band 600 MHz - 60 GHz. In addition, as previously discussed, this dispersion is accompanied by an localized rise in the corresponding imaginary permittivity $\epsilon''(f)$ with peak occurring at the relaxation frequency $f_e = 1/2\pi\tau_e$. As indicated, the corresponding peak amplitude is $(\epsilon_{dc} - \epsilon_{\infty})/2$. Note that the Debye permittivity spectrum illustrated in Figure 16a is perfectly consistent with the equivalent circuit and associated conductivity dispersion depicted in Figure 12.

A common generalization of the Debye spectrum is (Cole and Cole, 1941)

$$\epsilon(f) = \epsilon_{\infty} + \frac{\epsilon_{dc} - \epsilon_{\infty}}{1 + (i2\pi f\tau_e)^{1-c_1}}, \quad (38)$$

or

$$\epsilon'(f) = \epsilon_{\infty} + \frac{(\epsilon_{dc} - \epsilon_{\infty}) [1 + (2\pi f\tau_e)^{1-c_1} \sin(c_1\pi/2)]}{1 + 2(2\pi f\tau_e)^{1-c_1} \sin(c_1\pi/2) + (2\pi f\tau_e)^{2(1-c_1)}}$$

⁵ The characteristic Debye relaxation time (von Hippel, 1984) is $\tau_e = 3V\eta/kT$, where V represents an equivalent spherical molecular volume, η denotes effective fluid viscosity, k is Boltzmann's constant and T denotes temperature. For water at 25° C (298° K), with $\eta=0.01$ poise and $V \approx 2.0 \times 10^{-10} \text{ m}^3$, $\tau_e \approx 2.5 \times 10^{-11} \text{ s}$, implying a relaxation frequency $\nu_e \approx 4.0 \times 10^{10} \text{ Hz}$.

⁶ Note the similarity between equations (36-37) and equations (30-31) describing the relaxation of free ions in solution.

$$\epsilon''(f) = \frac{(\epsilon_{dc} - \epsilon_{\infty}) [(2\pi f\tau_e)^{1-c_e} \cos(c_e\pi/2)]}{1 + 2(2\pi f\tau_e)^{1-c_e} \sin(c_e\pi/2) + (2\pi f\tau_e)^{2(1-c_e)}},$$

where the so-called Cole-Cole distribution parameter $0.0 \leq c_e \leq 1.0$, controls the effective width of the dispersion band surrounding the relaxation frequency as depicted in Figure 16b. The so-called Cole-Cole spectrum is a more generally applicable model for the dielectric dispersion of polar substances. Note that the Debye relation arises for $c_e = 0$. Figure 16c displays the corresponding relationship between real ϵ' and imaginary ϵ'' components of the complex permittivity in Cole-Cole format ⁷.

Alternatively, as per equation (22) for the complex electrical conductivity, the Debye relation can be generalized as

$$\epsilon(f) = \epsilon_{\infty} \left[1 + \zeta \int_0^{\infty} \frac{i2\pi f\tau T_e(\tau)}{1 + i2\pi f\tau} d\tau \right] \quad (39)$$

for a continuous time constant distribution T_e , where $\zeta = (\epsilon_{dc} - \epsilon_{\infty})/\epsilon_{\infty}$ defines the dielectric chargeability. Here, the Debye spectrum arises for $T_e(\tau) = \delta(\tau - \tau_e)$.

Like the electrical conductivity, the electrical permittivity of a soil depends principally upon moisture content. In contrast, however, the electrical permittivity is relatively insensitive to the ionic strength of the soil water (Olhoeft, 1985; Schön, 1996). Consequently, to first approximation soil electrical permittivity is predicted by a volume weighted average of constituent permittivities. In particular, the CRIM (complex refractive index method) relation

$$\sqrt{\epsilon(f)} = \phi \sqrt{\epsilon_w(f)} + (1 - \phi) \sqrt{\epsilon_s(f)} \quad (40)$$

predicts the bulk electrical permittivity of a saturated soil as the porosity ϕ -weighted average of water $\epsilon_w(f)$ and solid $\epsilon_s(f)$ phase permittivities. The equivalent relation for a partially saturated, three phase soil is

$$\sqrt{\epsilon(f)} = \phi S_w \sqrt{\epsilon_w(f)} + (1 - \phi) \sqrt{\epsilon_s(f)} + \phi(1 - S_w) \sqrt{\epsilon_a(f)} \quad (41)$$

where S_w denotes partial saturation and $\epsilon_a(f) \approx \epsilon_0$ represents the electrical permittivity of entrained air. Foregoing relations are depicted by solid curves in Figures 17a, depicting the predicted dielectric constant as a function of partial saturation for a soil having porosity $\phi = 0.4$. The bounding curves are derived via an effective medium theory formulation known as the Bruggeman-Hanai-Sen (BHS) relation, accounting

⁷ For the particular case of a Debye spectrum ($c_e = 0$), the relation is semicircular with centre at $\epsilon' = (\epsilon_{dc} + \epsilon_{\infty})/2$, $\epsilon'' = 0$ and radius $(\epsilon_{dc} - \epsilon_{\infty})/2$. For $c_e > 0$, the centre is depressed beneath the real permittivity axis such that a radial line segment joining the centre and $\epsilon' = \epsilon_{\infty}$, $\epsilon'' = 0$ forms an interior angle of $\theta = c_e\pi/2$ with the real permittivity axis.

for structural characteristics of the soil as well as porosity and partial saturation. The fundamental relation is

$$\epsilon = \epsilon_w \phi^m \left[\left(1 - \frac{\epsilon_s}{\epsilon_w} \right) / \left(1 - \frac{\epsilon_s}{\epsilon} \right) \right]^m = \epsilon_w \phi^m \left(\frac{\epsilon_w - \epsilon_s}{\epsilon - \epsilon_s} \right)^m \left(\frac{\epsilon}{\epsilon_w} \right)^m, \quad (42)$$

where the so-called cementation index m , a function of grain geometry, varies between $m = 1.5$ for unconsolidated spherical grains, and $m = 2.0$ for highly consolidated-cemented oblate grains⁸. In contrast with the CRIM model, the BHS relation implicitly assumes pore-space interconnectivity. Bulk electrical permittivity for a three phase medium is obtained by consecutive applications of the foregoing relation, initially for the permittivity of the pore water/air (fluid) mixture and subsequently for the bulk, fluid saturated soil. We note that numerical methods are required for solution of BHS equation (Wobschall, 1977).

Figure 17b compares CRIM and BHS relations for three phase soils with an empirical relation derived by Topp *et al.* (1980), relating measured dielectric constant to a cubic polynomial with volumetric water content $\psi_w = S_w \phi = V_w/V_t$ as the independent variable

$$\kappa(\psi_w) = 3.03 + 9.3\psi_w + 146.0\psi_w^2 - 76.7\psi_w^3. \quad (43)$$

Here, V_w and V_t denote water volume and total soil volume, respectively. The Topp equation was derived via data acquired by the time-domain relectometry (TDR) method⁹ on a broad range of soils having widely variable textures, porosities and partial saturations. Although there is generally better correspondance with the CRIM model, comparison with BHS relations for $c=1.5$ and $c=2.0$ suggests of a saturation dependent transition in the effective distribution of soil water within the pore volume (Greaves, *et al.*, 1996).

In general, permittivity increases in direct proportion to partial saturation. However, a combination of experimental and theoretical studies have demonstrated that the distribution and state of water within the soil pore volume also has an influence on the electrical permittivity of the bulk soil. Laboratory measurements (Knight and Knoll, 1990) have established a pronounced hysteresis of the real permittivity as depicted in Figure 18a. Above a transitional saturation S_w^* , associated with surface adsorption, the electrical permittivity at a given partial saturation is substantially

⁸ Note the direct connection to Archie's law for electrical conductivity. On writing the equivalent BHS relation for electrical conductivity and assuming non-conductive matrix grains $\sigma_s = 0$, one obtains $\sigma = \sigma_w \phi^m$, as per equation (19) with $a = 1$, $S_w = 1$.

⁹ The standard time domain reflectometry (TDR) method involves measuring the propagation velocity v of a line-guided microwave pulse within the material under test. A corresponding dielectric constant is estimated via the relation $\tilde{\kappa} = v^2/v_0^2$, assuming a non-conductive ($\tilde{\sigma}/2\pi f\tilde{\epsilon} \ll 0.01$) medium ($v_0 = 1/\sqrt{\mu_0\epsilon_0}$ denotes the electromagnetic velocity of free-space).

higher during wetting (sorption) than during drainage (desorption). In connection with hysteresis in the relation between water content and soil suction (Figure 4c), the permittivity offset is attributed to a relative redistribution of liquid and gas phases within the pore volume. Numerical modelling (Endres and Knight, 1992) demonstrates that both pore geometry and the configuration of fluid within the pore space can have a significant influence on the resulting electrical permittivity ¹⁰.

In general, below transitional saturation, soil water is principally adsorbed water, tightly bound to soil matrix surfaces by electrostatic forces. As a result, the orientational polarizability of adsorbed water molecules is substantially lower than that of capillary or free water. Thus, an incremental increase in partial saturation below S_w^* produces a relatively small increase in soil electrical permittivity, compared with an equivalent increment above S_w^* . Experimental data displayed in Figure 19 illustrate this effect. In addition, Saarenketo (1998) has identified a similar contrast between the polarizability of capillary (viscous) and free water.

Comparison of complex permittivity spectra for bulk water and partially saturated soils (Hoekstra and Delaney, 1974) demonstrates that the effective relaxation frequency is relatively lower and the dispersive region narrower for water in soils, compared with water in bulk. In particular, observed permittivity dispersion was closely modelled by superposition of Cole-Cole spectra for two separate closely spaced relaxations, the higher frequency relaxation corresponding closely with that of bulk water. Although the lower frequency relaxation is consistent with an effectively more viscous adsorbed phase, Hoekstra and Delaney (1974) were explicit that such a connection was not supported by evidence that spectra were identical for sand and clay soils.

A survey of proposed dielectric mixing relations by Wang and Schmugge (1980) established that none of the relations, including CRIM and BHS, adequately model the adsorption effect. An alternative empirical model was developed, treating adsorbed water as a separate phase, having hybrid permittivity characteristics between those of bulk water and ice (Hasted, 1973; DeLoor, 1983). The complex permittivity is defined in relation to transitional water content $\psi_w^* = \phi S_w^*$ as

$$\epsilon(f) = \psi_w \epsilon_x(f) + (1 - \phi) \epsilon_s(f) + (\phi - \psi_w) \epsilon_a(f), \quad \psi_w \leq \psi_w^* \quad (44)$$

$$\epsilon_x(f) = \epsilon_i(f) + [\epsilon_w(f) - \epsilon_i(f)] \frac{\psi_w}{\psi_w^*} \Gamma$$

and

$$\epsilon(f) = \psi_w \epsilon_x(f) + (\psi_w - \psi_w^*) \epsilon_w(f) + (1 - \phi) \epsilon_s(f) + (\phi - \psi_w) \epsilon_a(f), \quad \psi_w > \psi_w^* \quad (45)$$

¹⁰ As per Figure 18b, electrical conductivity exhibits a similar saturation hysteresis with higher conductivity observed during wetting (sorption) than during drying (desorption). Again, hysteresis is attributed to changing fluid configuration within the pore volume as related to saturation dependent matric potential.

$$\epsilon_x(f) = \epsilon_i(f) + [\epsilon_w(f) - \epsilon_i(f)] \Gamma$$

where ϕ denotes porosity and ϵ_x represents the permittivity of adsorbed water, with permittivities ϵ_a , ϵ_w , ϵ_s and ϵ_i for air, water, matrix solids and ice, respectively. Γ is an adjustable regression parameter. On the basis of over twenty soils, having a wide range of textures, a strongly correlated linear relation

$$\psi_w^* = 0.49\psi_w + 0.165 \quad (46)$$

has been established between transitional water content ψ_w^* and wilting point ψ_w , defined in terms of soil texture as

$$\psi_w = 0.06774 - 0.00064F_s + 0.00478F_c, \quad (47)$$

where F_s and F_c denote dry weight fractions of sand and clay, respectively ¹¹. Finally, the imaginary permittivity predicted by equations (44-47) is augmented by addition of a loss factor $\epsilon''_\sigma = \zeta\psi_w^2$ where ζ is a texture dependent regression parameter, having representative values $\zeta=0.0$, 8.0 and 26.0 for sands, loams and clays, respectively ¹².

Combined, equations (44-47) constitute a well constrained model for estimating the complex electrical permittivity on the basis of soil texture ¹³. Figure 20 displays predicted saturation dependent electrical permittivity for three hypothetical soils (sand, loam and clay), assuming fixed porosity $\phi = 0.5$. CRIM and Topp relations are overlayed for comparison. Due to a relatively high proportion of adsorbed water, the clay soil possesses a relatively low real permittivity independent of partial saturation. In contrast, the sand, having relatively limited charged surface area, contains relatively little adsorbed water and, consequently, is characterized by a relatively high permittivity. A reverse scenario is observed for associated imaginary permittivities, where the loss factor ϵ''_σ is responsible for rapidly increasing permittivities for clay and loam soils in connection with enhanced surface related free-charge polarization phenomena. In practice, the relationship between soil texture and electrical permittivity is potentially more complicated. In particular, although the foregoing model predicts a relatively low real permittivity for the clay soil at any given moisture content, clays are typically characterized by a relatively high porosity (Figure 3c) and possess a significantly higher suction (Figure 4b) than coarser grained soils for a given

¹¹ Equation (47) was established via multiple regression of more than 100 soil samples (Schmugge *et al.*, 1976).

¹² According to Wang and Schmugge (1980) an adequate low-frequency approximation for conductive loss is $\epsilon''_\sigma = (v/f)\sigma_{dc}$, where v denotes the effective propagation velocity (m/ns) at frequency f (GHz) as discussed by DeLoor (1983).

¹³ An alternative strategy for modelling soil dielectric properties, combining multiphase dielectric mixing formulae and empirically derived soil texture-moisture permittivity relations is reported by Hallikainen *et al.* (1985), Dobson *et al.* (1985) and Peplinski *et al.* (1995).

partial saturation. Thus, clay soils can be characterized by a relatively high moisture content and, thus, higher dielectric constant, despite a greater proportion of adsorbed water.

Representative laboratory data for a range of saturated sand-clay mixtures are tabulated and displayed graphically in Figure 21a (Olhoeft, 1987). Note the coupled correlation between clay content, saturated porosity and real relative permittivity at 100 MHz. Higher clay content is associated with increased saturation water content and correspondingly higher permittivity. On the other hand, Figure 21b presents the frequency distribution of field-measured real permittivities for clay, loam and sand soils as reported by Daniels (1997). Here, clays are predominantly associated with relatively low real permittivities as per the findings of Wang and Schmutge (1980). In short, although clays possess the potential for higher water content and a correspondingly higher permittivity, the natural moisture level of clay soils is ordinarily well below saturation and typically such that a majority resides in an adsorbed or bound state. Consequently, loam soils, which possess maximum available water (Figure 5), also retain on average a relatively high capillary water content and, thus, are typically characterized by a comparatively higher real electrical permittivity.

In summary, like the electrical conductivity, the electrical permittivity of soils is predominantly determined by water content. In contrast, however, the salinity of soil water has relatively little influence. Bulk soil permittivity is obtained as a direct combination of constituent permittivities via a range of mixing formulas. Again, surface area effects and, consequently, texture are critical factors. In particular, surface adsorbed or hygroscopic water is characterized by a relatively lower electrical polarizability than bulk water. Consequently, for a given volumetric water content, soils possessing greater charged surface area, particularly clays, are characterized by a relatively high proportion of hygroscopic water and, thus, a lower real electrical permittivity. In part, however, this trend is compensated by an associated propensity of fine textured soil to attract and retain water.

Also like the electrical conductivity, the electrical permittivity is dispersive and, thus, complex valued. However, while dispersion of the electrical conductivity is due to interfacial polarization of aqueous free-charge at frequencies below 100 MHz, electrical permittivity dispersion is principally due to bound-charge orientational polarization of the water molecule in the GHz region. This is particularly evident on observing the variation in soil dielectric constant as a function of temperature near 0°C. Figure 22a displays the temperature dependent complex permittivity for a silt soil with gravimetric moisture contents of 5.0, 10.0 and 15.0 percent. Samples were initially frozen at -20°C. At temperatures substantially below 0°C water molecules are rigidly bound in solid crystalline structure and, consequently, are not polarizable by via rotation. However, as soil water changes state at 0°C molecules become free to rotate and a marked rise in permittivity is observed in connection with resulting orientational polarization. The effect is clearly moisture dependent with relatively little

effect observable at 5.0 percent moisture. Transitional characteristics are attributable to the existence of residual liquid water below 0°C and adsorption related viscosity effects above 0°C (Hoekstra and Deianey, 1974). Corresponding data for a clay soil in Figure 22b exhibit a similar pattern.

Representative empirical data on the electrical permittivity of soils are tabulated in Appendix I.

2.3 MAGNETIC PERMEABILITY

As described by the constitutive relation $B = \mu H$, magnetic permeability μ relates magnetic field intensity H to the resulting flux density B induced within a given medium. In effect, the permeability quantifies the medium's susceptibility to magnetization (magnetic polarization), according to the relation $M = kH$, where the related constant $k = (\mu/\mu_0) - 1$ is termed magnetic susceptibility. Here, $\mu_0 = 4\pi \times 10^{-7}$ H/m is the magnetic permeability of free space.

As for electrical polarization, several independent mechanisms are responsible for net magnetization. Principally, these include diamagnetism, paramagnetism and ferromagnetism¹⁴. Diamagnetism arises in connection with the orbital motion of electrons about the nucleus of an atom, producing a net magnetic dipole moment in opposition to the net ambient field. The associated magnetic susceptibility is, thus, negative valued.

Individual electrons also possess an intrinsic spin magnetic moment. Although the electronic structure of an atom is generally such that electrons having opposing spin moments are paired and, thus, cancel, a given atom or molecule can possess several uncoupled electrons, giving rise to a substantial net magnetic moment. Partial alignment of residual atomic magnetic moments with an applied external field results in paramagnetism. The associated positive valued magnetic susceptibility depends on the specific electronic configuration. However, in general, paramagnetic susceptibility is ordinarily orders of magnitude greater than underlying diamagnetism and, consequently, only those substances characterized by perfect cancellation of spin magnetic moments are described as diamagnetic.

Paramagnetism of certain solid substances, including iron, is such that residual spin magnetic moments within a given localized region, referred to as a magnetic domain, are mutually oriented (Figure 23a). Moreover, on application of an external magnetic field, domains with magnetization parallel to the applied field are preferentially enlarged at the expense of neighboring domains, resulting in a net ferromagnetism parallel to the applied field as indicated in Figure 23b.¹⁵ The corresponding

¹⁴ For a detailed discussion of magnetic polarization mechanisms see Purcell (1963) and Strangway (1967).

¹⁵ The origin and organization of magnetic domains involves complicated internal energy considerations as described by Nagata (1961) and Strangway (1967).

magnetic susceptibility is typically several orders of magnitude greater than that for a purely paramagnetic material. In a related form of spontaneous magnetization, termed ferrimagnetism (Figure 23c), paramagnetic moments within a given domain are arranged such that moments on alternating sublattices are oriented parallel and antiparallel, with partial cancellation yielding diminished domain magnetization. Antiferromagnetism refers to the specific case of zero net domain magnetization.

As for electrical polarization processes, magnetic polarization arising on application of an external magnetic field decays on removal of the polarizing field with associated relaxation leading to dispersion of the frequency dependent permeability. It follows that the magnetic permeability, like its electrical counterparts, is a complex valued quantity

$$\mu(f) = \mu'(f) - i\mu''(f), \quad (49)$$

with in-phase $\mu'(f)$ and quadrature $\mu''(f)$ components related via analogous Kramers-Kronig relations

$$\mu''(f) = \frac{1}{\pi} \int_{-\infty}^{\infty} \frac{\mu'(\varphi)}{\varphi - f} d\varphi = -\frac{2f}{\pi} \int_0^{\infty} \frac{\mu'(\varphi)}{f^2 - \varphi^2} d\varphi \quad (50)$$

$$\mu'(f) = -\frac{1}{\pi} \int_{-\infty}^{\infty} \frac{\mu''(\varphi)}{\varphi - f} d\varphi = \frac{2}{\pi} \int_0^{\infty} \frac{\varphi\mu''(\varphi)}{f^2 - \varphi^2} d\varphi. \quad (51)$$

Similar relations are satisfied by the magnetic susceptibility.

Moreover, magnetic polarization and related relaxation phenomena are also effectively modelled via Debye or Cole-Cole dispersion relations. In particular, in analogy with equations (22) and (39), the equivalent generalized expression for magnetic permeability is

$$\mu(f) = \mu_{\infty} \left[1 + \xi \int_0^{\infty} \frac{i2\pi f \tau T_{\mu}(\tau)}{1 + i2\pi f \tau} d\tau \right], \quad (52)$$

where $\xi = (\mu_{dc} - \mu_{\infty})/\mu_{\infty}$ defines an analogous magnetic chargeability, in terms of static μ_{dc} and high frequency μ_{∞} permeabilities, and $T_{\mu}(\tau)$ denotes a continuous time constant distribution for constituent magnetic relaxation processes.

For $T_{\mu}(\tau) = \delta(\tau - \tau_{\mu})$, equation (52) yields the standard Debye spectrum

$$\mu(f) = \mu_{\infty} + \frac{\mu_{dc} - \mu_{\infty}}{1 + i2\pi f \tau_{\mu}}, \quad (53)$$

or

$$\mu'(f) = \mu_{\infty} + \frac{\mu_{dc} - \mu_{\infty}}{1 + (2\pi f \tau_{\mu})^2} \quad (54)$$

$$\mu''(f) = \frac{2\pi f \tau_{\mu} (\mu_{dc} - \mu_{\infty})}{1 + (2\pi f \tau_{\mu})^2}. \quad (55)$$

The corresponding Cole-Cole spectrum is

$$\mu(f) = \mu_{\infty} + \frac{\mu_{dc} - \mu_{\infty}}{1 + (i2\pi f\tau_{\mu})^{1-c_{\mu}}} \quad (56)$$

or

$$\mu'(f) = \mu_{\infty} + \frac{(\mu_{dc} - \mu_{\infty}) [1 + (2\pi f\tau_{\mu})^{1-c_{\mu}} \sin(c_{\mu}\pi/2)]}{1 + 2(2\pi f\tau_{\mu})^{1-c_{\mu}} \sin(c_{\mu}\pi/2) + (2\pi f\tau_{\mu})^{2(1-c_{\mu})}} \quad (57)$$

$$\mu''(f) = \frac{(\mu_{dc} - \mu_{\infty}) [(2\pi f\tau_{\mu})^{1-c_{\mu}} \cos(c_{\mu}\pi/2)]}{1 + 2(2\pi f\tau_{\mu})^{1-c_{\mu}} \sin(c_{\mu}\pi/2) + (2\pi f\tau_{\mu})^{2(1-c_{\mu})}}, \quad (58)$$

for $0 < c_{\mu} < 1$. Figure 23c displays normalized Cole-Cole permeability spectra for a range of c_{μ} . As previously discussed in connection with the electrical permittivity, the so-called distribution parameter c_{μ} controls the effective width of the dispersion band surrounding the relaxation frequency. Again, the Debye spectrum, obtained for $c_{\mu} = 0$ is associated with minimum dispersion bandwidth.

A laboratory study of single crystal magnetite (Galt, 1952) confirms the general validity of the foregoing dispersion relations for description of ferromagnetic polarization. In particular, Galt (1952) attributed magnetic dispersion at approximately 20 kHz (Figure 24a) ($\tau_{\mu} = 5.0 \times 10^{-5}$ s) to domain wall relaxation. The associated dispersion parameter characterizing the extent of frequency dependent domain wall mobility is approximately $c_{\mu} = 0.63$. Similar measurements on polycrystalline samples were reported by Olhoeft and Strangway (1974). In general, observed relaxation frequencies are lower, below 550 Hz, and the associated dispersion band narrower ($c_{\mu} < 0.15$). More recently, Olhoeft and Capron (1994) reported results of a laboratory study of high frequency magnetic dispersion in granular media. Graduated mixtures comprising between 10 and 100 percent by weight fine grained iron filings in dry silica powder were analyzed, with associated dispersion characterized via "best-fit" Cole-Cole parameters. Figure 24b displays measured real relative magnetic permeability $\mu'_r = \mu'/\mu_0$ and magnetic loss $\tan \delta_{\mu} = \mu''/\mu'$ as functions of frequency. Low frequency relative permeability ranged from approximately $\mu_{dc}/\mu_0 = 1.15 - 10.95$ with increasing iron content. Corresponding relaxation frequencies were approximately $f_{\mu} = 1/\tau_{\mu} = 3.3$ GHz-715 MHz, with associated distribution parameters relatively consistent about mean value $c_{\mu} = 0.27$.

In contrast with the electrical properties of soils which are predominantly controlled by water content and relatively insensitive to the mineralogy of matrix constituents, soil magnetic permeability is effectively uninfluenced by soil moisture and almost entirely dependent on matrix mineralogy. In particular, the principal source of soil magnetism is iron, residing in one form or another, within residual parent material, secondary mineralization and aqueous phase chemical species. Chemical weathering of rocks and soils constitutes a continuous source of aqueous ions that are either leached from the soil or react to form a range of quasi-stable biproducts

residing within the soil matrix. The most common biproducts are clay minerals and hydrous oxides of aluminium and iron.

Weathering of iron-bearing minerals typically releases the ferrous ion Fe^{2+} into the soil water where subsequent reaction depends upon local pH-Eh conditions of the soil environment as summarized in Figure 25. In particular, where Fe^{2+} is released via hydrolysis within an oxidizing environment, it is oxidized to the ferric state Fe^{3+} upon interaction with soil water and rapidly precipitated as a ferric oxide or hydrous oxide of iron over the normal range of soil pH (Krauskopf, 1967; Birkeland, 1974; Bohn *et al.*, 1985). In contrast, Fe^{2+} released within a reducing environment remains in the ferrous state with significantly enhanced ionic mobility and, thus, remains a relatively free ion within the soil water. Eh-pH diagrams charting effective stability fields for various iron ions and compounds under a wide range of specified equilibrium conditions are described by Garrels and Christ (1965).

Where precipitated under aerated soil conditions, ferric oxides and hydroxides typically occur as amorphous gels or discrete particles, coating larger silicate grains that constitute the bulk soil matrix. It is these relatively abundant compounds that impart the characteristic hues of iron rich soils. In addition, common iron oxides, including hematite $\alpha\text{Fe}_2\text{O}_3$ and its hydration product goethite FeOOH , are also weakly magnetic and account for minor soil magnetization¹⁶. However, soils possessing substantial magnetic susceptibility inevitably incorporate significant concentrations of the ferrimagnetic oxides, principally magnetite Fe_3O_4 and maghaemite $\gamma\text{Fe}_2\text{O}_3$.

Magnetite is occasionally observed as a secondary mineral under strongly reducing, alkaline conditions, however its occurrence in near surface soils is largely as residual mineral grains, inherited via weathering of igneous and metamorphic parent material. In contrast, maghaemite is predominantly a secondary compound, ultimately arising via oxidation of magnetite. In particular, alternating exposure to aerobic and anaerobic environments promotes the conversion of constituent iron oxide from relatively abundant and weakly ferromagnetic haematite to ferrimagnetic magnetite, via reduction, and on subsequent re-oxidation to maghaemite, having a magnetic susceptibility approximately 200 times that of hematite (Le Borgne, 1955; Mullins, 1977). This so-called magnetic conversion occurs as result of repeated burning of overlying vegetation or via microbial activity in connection with alternating saturation and drying of more or less organic (humic) soils, particularly under warm and humid (tropical) climatic conditions (Tucker, 1952; Cook and Carts, 1962).

Finally, it is emphasized that the bulk of soil magnetic susceptibility is due to sub-clay size particles. In particular, ferrimagnetic grains having dimensions in the range 0.5×10^{-4} mm - 1.0×10^{-4} mm are typically single domain and, consequently, possess relatively higher magnetic susceptibility than larger, multidomain grains of

¹⁶ Hematite $\alpha\text{Fe}_2\text{O}_3$ is effectively antiferromagnetic, retaining only weak parasitic or imperfect ferromagnetism as indicated in Figure 23c. Goethite or limonite FeOOH is weakly paramagnetic.

the same material. In addition, the susceptibility of such grains increases with time due to magnetic viscosity (Néel, 1955). Although magnetic domains do not arise in grains smaller than approximately 0.5×10^{-4} mm these grains also possess enhanced magnetic susceptibility due to a similar phenomenon referred to as superparamagnetism. Both effects lead to significant magnetic dispersion (Dabas *et al.*, 1992).

Cook and Carts (1962) reported a survey of soil magnetic properties for more than 250 samples spanning a representative range of soil types. Although the bulk of susceptibility was with few exceptions attributed to residual magnetite grains, results presented in Figure 26 display only a weak correlation with parent material. Average magnetic susceptibilities compiled by Telford *et al.* (1976) for several parent materials are overlayed for comparison. Of particular interest is the relatively high magnetic susceptibility of soils formed on low iron content parent materials. In general, higher magnetic susceptibilities were observed for Panamanian soils and attributed to organic reduction of limonite (magnetic conversion) under relatively warm and moist tropical climate conditions. Additional empirical data on the magnetic susceptibility of soils are tabulated in Appendix I.

In practice, it is commonly assumed that magnetic dispersion is negligible $\mu(f) = \mu'(f)$ ($\mu''(f) = 0$). It is also ordinarily asserted that the magnetic permeability of soils does not deviate appreciably from the free space value $\mu_0 = 4\pi \times 10^{-7}$ H/m. In fact, although the relative magnetic permeability $\mu_r = \mu'(0)/\mu_0 = k + 1$ possesses a maximum of approximately $\mu_r = 15$ for pure magnetite, soils seldom possess relative permeabilities exceeding $\mu_r = 1.1$. For instance, the maximum susceptibility of $k = 1.0 \times 10^{-2}$ emu¹⁷ displayed in Figure 26, translates to a relative permeability of approximately $\mu_r = 1.01$, a one percent deviation from free space permeability. Nonetheless, because magnetic permeability scales the remainder of the propagation constant as per equation (15), even moderate susceptibility variation can give rise to a significant electromagnetic signature. Indeed, the Slingram electromagnetic method is employed routinely to map patterned variability in soil magnetic susceptibility in connection with archaeological site assessment (Tabbagh, 1985). Moreover, the previously described findings reported by Olhoeft and Capron (1994) demonstrate the potential significance of magnetic dispersion at radar frequencies. Consequently, while the simplifying assumption $\mu_r = 1$ is commonly justified, it is important to appreciate the potential existence of "soil noise", arising in connection with spatial variation and intrinsic dispersion of the magnetic permeability. In general, magnetic effects deserve due consideration in respect of specific soil conditions.

3.0 EFFECTIVE MATERIAL PARAMETERS

As per foregoing sections 2.1-2.3, soil electromagnetic properties are, in general, dis-

¹⁷ To convert susceptibility in emu to SI or so-called rationalized susceptibility, multiply by 4π .

persive and, consequently, complex valued quantities

$$\sigma = \sigma' + i\sigma'' \quad (23)$$

$$\epsilon = \epsilon' - i\epsilon'', \quad (32)$$

$$\mu = \mu' - i\mu'' \quad (49)$$

Accordingly, substitution in equation (15) leads to the following explicit expression for the propagation constant.

$$\gamma = 2\pi f (\mu' + i\mu'')^{1/2} \left[i \frac{(\sigma' + i\sigma'')}{2\pi f} - (\epsilon' - i\epsilon'') \right]^{1/2}. \quad (59)$$

Now, on rearranging terms, we obtain the practical form

$$\gamma = 2\pi f (\mu' + i\mu'')^{1/2} \left[\frac{i}{2\pi f} (\sigma' + 2\pi f\epsilon'') - \left(\epsilon' + \frac{\sigma''}{2\pi f} \right) \right]^{1/2}, \quad (60)$$

$$= 2\pi f \bar{\mu}^{1/2} \bar{\epsilon}^{1/2} \left(i \frac{\bar{\sigma}}{2\pi f \bar{\epsilon}} - 1 \right)^{1/2}, \quad (61)$$

defining real-valued effective conductivity

$$\bar{\sigma} = \sigma' + 2\pi f\epsilon'' \quad (62)$$

and real-valued effective permittivity ¹⁸

$$\bar{\epsilon} = \epsilon' + \frac{\sigma''}{2\pi f}. \quad (63)$$

In addition, for the time being, we shall neglect magnetic losses, asserting the common assumption $\bar{\mu}(f) = \mu'(f)$.

Although it is recognized that equations (15) and (61) have precisely the same form, practical justification for introducing effective electrical parameters is perhaps not immediately obvious. To appreciate the significance of effective parameters, we return to Ampere's Law. From equation (10), we obtain

$$\begin{aligned} \nabla \times \mathbf{H} &= [(\sigma' + i\sigma'') + i2\pi f(\epsilon' - i\epsilon'')] \mathbf{E} \\ &= \left[(\sigma' + 2\pi f\epsilon'') + i2\pi f \left(\epsilon' - \frac{\sigma''}{2\pi f} \right) \right] \mathbf{E} \end{aligned}$$

¹⁸ Note that sign conventions adopted in (23) and (32) are consistent with positive-valued effective parameters $\bar{\sigma}$ and $\bar{\epsilon}$ and physically realizable attenuation rate.

$$= (\bar{\sigma} + i2\pi f\bar{\epsilon}) \mathbf{E}. \quad (64)$$

It follows that while it is practically impossible to distinguish between free-charge and bound-charge current density contributions, net in-phase and quadrature components of the current density $\mathbf{J} = \bar{\mathbf{J}}_c + i\bar{\mathbf{J}}_d$, as directly related to effective electrical parameters $\bar{\mathbf{J}}_c = \bar{\sigma}\mathbf{E}$ and $\bar{\mathbf{J}}_d = 2\pi f\bar{\epsilon}$, are readily separable. Consequently, although the fact is seldom emphasized or perhaps appreciated, it is these effective electrical parameters that are measured in the laboratory and cited in the scientific literature ¹⁹.

In particular, Figure 27 depicts measured frequency dependent effective electrical parameters for a clay-loam soil as compiled from a number of independent studies. Note that low-frequency effective permittivity displayed in Figure 27a far exceeds the corresponding value for water due to the influence of free charge polarization processes as quantified by the quadrature conductivity term $\sigma''(f)/2\pi f$ (Figure 12b). Similarly, the high-frequency effective conductivity is dominated by water relaxation effects described by the quadrature permittivity term $2\pi f\epsilon''(f)$ (Figure 16a).

The composite nature of effective parameters is emphasized in Figure 27b, where complex parameters are modelled via Debye dispersion relations for typical soil values $\sigma_{dc} = 5 \text{ mS/m}$, $\sigma_{\infty} = 10 \text{ mS/m}$, $\tau_{\sigma} = 0.01 \text{ s}$, $\epsilon_{dc} = 15\epsilon_0$, $\epsilon_{\infty} = 5\epsilon_0$ and $\tau_{\epsilon} = 1.0 \times 10^{-10} \text{ s}$. Note the relative magnitude of constituent parameters, particularly the predominance of quadrature terms $2\pi f\epsilon''$ and $\sigma''/2\pi f$, due to frequency scaling. In general, the relative influence of contributing parameters is soil dependent.

Figure 28a displays the corresponding loss tangent

$$\tan \delta = \frac{\bar{\sigma}}{2\pi f\bar{\epsilon}} = \frac{\sigma' + 2\pi f\epsilon''}{2\pi f\epsilon' + \sigma''}, \quad (68)$$

given by the ratio of quadrature and in-phase components of the effective current density (Figure 28b) ²⁰. In effect, the loss tangent is a relative measure of the energy dissipation characteristics of the medium. An increase in the loss tangent indicates that a greater proportion of propagating energy is dissipated through ohmic heating and radiation damping compared with energy stored via combined free and bound charge polarization. Figure 29 illustrates the cumulative attenuation effect of free-charge conduction, orientational relaxation of constituent H_2O , scattering and colloidal interface relaxation.

To appreciate the practical significance of the loss tangent, we return to the propagation constant to further investigate the roles of effective electrical parameters

¹⁹ An alternative parameter formulation in connection with high frequency electrical permittivity measurements defines complex effective permittivity as $\bar{\epsilon} = \bar{\epsilon}' + i\bar{\epsilon}''$, with $\bar{\epsilon}' = \epsilon' + \sigma''/2\pi f$ and $\bar{\epsilon}'' = \epsilon'' + \sigma'/2\pi f$. Similarly, one could define complex electrical conductivity as $\bar{\sigma} = \bar{\sigma}' + i\bar{\sigma}''$ with $\bar{\sigma}' = \sigma' + 2\pi f\epsilon''$ and $\bar{\sigma}'' = \sigma'' + 2\pi f\epsilon'$.

²⁰ Note that alternative effective parameter definitions¹⁹ yield equivalent expressions for the loss tangent $\tan \delta = \bar{\epsilon}''/\bar{\epsilon}' = \bar{\sigma}''/\bar{\sigma}' = \bar{\sigma}/2\pi f\bar{\epsilon}$.

in controlling the propagation and attenuation of electromagnetic waves within a material medium.

Recall the governing equations

$$\nabla^2 \mathbf{E} = \gamma^2 \mathbf{E} \quad (13)$$

$$\nabla^2 \mathbf{H} = \gamma^2 \mathbf{H} \quad (14)$$

and associated propagation constant

$$\gamma = 2\pi f \bar{\mu}^{1/2} \bar{\epsilon}^{1/2} \left(i \frac{\bar{\sigma}}{2\pi f \bar{\epsilon}} - 1 \right)^{1/2}. \quad (61)$$

For a non-conductive medium we have $\tan \delta = \bar{\sigma}/2\pi f \bar{\epsilon} \ll 1$ and equation (61) yields the approximation

$$\gamma \approx -2\pi f (\bar{\mu} \bar{\epsilon})^{1/2}. \quad (69)$$

On substitution in the governing equations (13) and (14) we obtain

$$\nabla^2 \mathbf{E} = \gamma^2 \mathbf{E} = (i2\pi f)^2 \bar{\mu} \bar{\epsilon} \mathbf{E} \quad (70)$$

$$\nabla^2 \mathbf{B} = \gamma^2 \mathbf{B} = (i2\pi f)^2 \bar{\mu} \bar{\epsilon} \mathbf{B} \quad (71)$$

or, via the derivative theorem of the Fourier transform, the equivalent time-domain wave equations

$$\nabla^2 \mathbf{E} = \bar{\mu} \bar{\epsilon} \frac{\partial^2 \mathbf{E}}{\partial t^2} \quad (72)$$

$$\nabla^2 \mathbf{B} = \bar{\mu} \bar{\epsilon} \frac{\partial^2 \mathbf{B}}{\partial t^2}, \quad (73)$$

describing attenuation free propagation at velocity $v = (1/\bar{\mu} \bar{\epsilon})^{1/2}$. In contrast, for a conductive soil $\tan \delta = \bar{\sigma}/2\pi f \bar{\epsilon} \gg 1$, equation (61) yields the approximation

$$\gamma \approx (i2\pi f \bar{\mu} \bar{\sigma})^{1/2}, \quad (74)$$

with substitution in equations (13) and (14) giving

$$\nabla^2 \mathbf{E} = \gamma^2 \mathbf{E} = i2\pi f \bar{\mu} \bar{\sigma} \mathbf{E} \quad (75)$$

$$\nabla^2 \mathbf{B} = \gamma^2 \mathbf{B} = i2\pi f \bar{\mu} \bar{\sigma} \mathbf{B}. \quad (76)$$

In contrast with wave equations arising in connection with a dielectric medium, the equivalent time-domain expressions

$$\nabla^2 \mathbf{E} = \bar{\mu} \bar{\sigma} \frac{\partial \mathbf{E}}{\partial t} \quad (77)$$

$$\nabla^2 \mathbf{B} = \bar{\mu}\bar{\sigma} \frac{\partial \mathbf{B}}{\partial t} \quad (78)$$

are diffusion equations, describing a fundamentally different mode of behaviour. In short, a conductive medium is incapable of propagating the electromagnetic disturbance due to rapid attenuation. More precisely, the polarized charge distribution relaxes more rapidly than the inducing field reverses polarity. The rate of diffusion into the soil is $v = (4\pi f)^{1/2} d = (4\pi f / \bar{\mu}\bar{\sigma})^{1/2}$, where $d = (1/\bar{\mu}\bar{\sigma})^{1/2}$ denotes the corresponding diffusivity.

Consequently, the loss tangent value provides a practical guide to the effective mode of electromagnetic interaction as illustrated in Figure 30. In particular, for $\tan \delta = \bar{\sigma} / 2\pi f \bar{\epsilon} < 0.01$, the soil is effectively dielectric, displacement currents (energy storage) dominate and wave propagation is the principal mode of interaction. In contrast, for $\tan \delta = \bar{\sigma} / 2\pi f \bar{\epsilon} > 100$, the soil is conductive, conduction currents (energy dissipation) dominate and diffusion is the primary mode of interaction. Finally, for intermediate values $0.01 < \tan \delta = \bar{\sigma} / 2\pi f \bar{\epsilon} < 100$, the soil is described as quasi-conductive, permitting attenuated wave propagation as per the full propagation constant. In particular, on substituting equation (61) in equations (13) and (14), we obtain

$$\nabla^2 \mathbf{E} = \gamma^2 \mathbf{E} = [(i2\pi f)^2 \bar{\mu}\bar{\epsilon} + i2\pi f \bar{\mu}\bar{\sigma}] \mathbf{E} \quad (79)$$

$$\nabla^2 \mathbf{B} = \gamma^2 \mathbf{B} = [(i2\pi f)^2 \bar{\mu}\bar{\epsilon} + i2\pi f \bar{\mu}\bar{\sigma}] \mathbf{B} \quad (80)$$

and the equivalent time domain equations

$$\nabla^2 \mathbf{E} = \bar{\mu}\bar{\epsilon} \frac{\partial^2 \mathbf{E}}{\partial t^2} + \bar{\mu}\bar{\sigma} \frac{\partial \mathbf{E}}{\partial t} \quad (81)$$

$$\nabla^2 \mathbf{B} = \bar{\mu}\bar{\epsilon} \frac{\partial^2 \mathbf{B}}{\partial t^2} + \bar{\mu}\bar{\sigma} \frac{\partial \mathbf{B}}{\partial t}, \quad (82)$$

including both propagation and diffusion terms.

Further insight is furnished by considering solutions of equations (13) and (14) having form

$$\mathbf{E} = \mathbf{E}_0 e^{i2\pi ft - \gamma z} = \mathbf{E}_0 e^{i2\pi ft - (\alpha + i\beta)z} \quad (83)$$

$$\mathbf{H} = \mathbf{H}_0 e^{i2\pi ft - \gamma z} = \mathbf{H}_0 e^{i2\pi ft - (\alpha + i\beta)z}, \quad (84)$$

where α and β denote in-phase and quadrature components of the propagation constant $\gamma = \alpha + i\beta$, respectively. In general, the foregoing solutions describe a plane electromagnetic disturbance traveling in the z -direction. Rewriting the solutions as

$$\mathbf{E} = \mathbf{E}_0 e^{-\alpha z} e^{i(2\pi ft - \beta z)} \quad (85)$$

$$\mathbf{H} = \mathbf{H}_0 e^{-\alpha z} e^{i(2\pi ft - \beta z)}, \quad (86)$$

we identify α as the effective attenuation rate and $2\pi f/\beta = v$ as the associated velocity.

Straight forward analysis of the propagation constant yields

$$\alpha = 2\pi f \left[\frac{\bar{\mu}\bar{\epsilon}}{2} \left(\left[1 + \left(\frac{\bar{\sigma}}{2\pi f\bar{\epsilon}} \right)^2 \right]^{1/2} - 1 \right) \right]^{1/2} \quad (87)$$

$$v = \frac{2\pi f}{\beta} = \left[\frac{\bar{\mu}\bar{\epsilon}}{2} \left(\left[1 + \left(\frac{\bar{\sigma}}{2\pi f\bar{\epsilon}} \right)^2 \right]^{1/2} + 1 \right) \right]^{-1/2} \quad (88)$$

Figure 31 displays theoretically predicted attenuation rate and velocity as a functions of frequency with conductivity and permittivity as constant parameters. Although dispersion of electrical parameters is ignored, two separate regimes are identified. A dispersive regime for $\tan \delta = \bar{\sigma}/2\pi f\bar{\epsilon} < 1$ and a non-dispersive regime for $\tan \delta = \bar{\sigma}/2\pi f\bar{\epsilon} > 1$. The transition occurs at frequency $f_T = \bar{\sigma}/2\pi\bar{\epsilon}$ ($\tan \delta = \bar{\sigma}/2\pi f\bar{\epsilon} = 1$). Moreover, we note that the influence of conductivity and permittivity on the effective velocity is restricted to dispersive and non-dispersive regimes, respectively.

Clearly, these observations are consistent with the foregoing designation of dielectric, quasi-conductive and conductive media as indicated by the associated loss tangent. Here, however, the controlling parameter is frequency. In particular, for a conductive medium, we have $\tan \delta = \bar{\sigma}/2\pi f\bar{\epsilon} \gg 1$, leading to dispersive relations $\alpha \approx (\pi f \bar{\mu} \bar{\sigma})^{1/2}$ and $v \approx (4\pi f / \bar{\mu} \bar{\sigma})^{1/2}$. As observed in Figure 31 for $f < f_T$, neither parameter is influenced by permittivity. Utilizing the complex identity $\sqrt{i} = (1+i)/\sqrt{2}$, the associated propagation constant follows as $\gamma = (1+i)(\pi f \bar{\mu} \bar{\sigma})^{1/2} = (i2\pi f \bar{\mu} \bar{\sigma})^{1/2}$ as per equation (74). Thus, it is emphasized that for given non-dispersive electrical parameters, a soil is effectively conductive for frequencies $f < f_T$ with associated electromagnetic interaction described by diffusion equations (75-78).

Similarly, for a dielectric medium, $\tan \delta = \bar{\sigma}/2\pi f\bar{\epsilon} \ll 1$, we obtain approximate relations ²¹ $\alpha \approx (\bar{\mu}\bar{\sigma}^2/4\bar{\epsilon})^{1/2}$ and $v = (1/\bar{\mu}\bar{\epsilon})^{1/2}$. As depicted in Figure 31 for $f > f_T$, both parameters are non-dispersive and velocity is independent of conductivity ²². However, comparing the associated propagation constant

$$\gamma \approx (\bar{\mu}\bar{\sigma}^2/4\bar{\epsilon})^{1/2} + i2\pi f(\bar{\mu}\bar{\epsilon}) \quad (89)$$

with equation (69), it is observed that despite non-dispersive behavior for $f > f_T$, associated electromagnetic interaction is not governed by equations (70-73) but, rather,

²¹ High frequency approximations are obtained via the binomial expansion $[1 + (\bar{\sigma}/2\pi f\bar{\epsilon})^2]^{1/2} \approx 1 + \frac{1}{2}(\bar{\sigma}/2\pi f\bar{\epsilon})^2$.

²² We note that relative insensitivity of TDR moisture measurements to soil conductivity variation requires that constituent frequencies comprising the transmitted pulse satisfy the dielectric condition $f \gg f_T = \bar{\sigma}/2\pi\bar{\epsilon}$.

by equations (79-82), describing attenuated wave propagation. Only for $\bar{\sigma} \approx 0$ is a medium strictly non-conductive and governed by pure wave equations. Independent of frequency, any soil possessing finite conductivity is also characterized by a finite attenuation rate as per Figure 31. Moreover, assuming nondispersive electrical parameters, the attenuation rate approaches a constant maximum value for $f > f_T$. Frequency dispersion effects are summarized in Figure 32.

The added effects of material dispersion are illustrated in Figure 33. Superimposed on velocity and attenuation spectra for constant conductivities, as depicted in Figure 31, are corresponding spectra for equivalent high frequency conductivities $\sigma_\infty = 1, 10, 100$ and 1000 mS/m, with chargeability $\eta = (\sigma_\infty - \sigma_{dc})/\sigma_\infty = 0.5$ and Debye relaxation constant $\tau_e = 0.001$ s ($f_e = 160$ Hz). Similarly, permittivity dispersion is modelled via a Debye spectrum with $\epsilon_{dc} = 15\epsilon_0$ H/m, $\epsilon_\infty = 5\epsilon_0$ H/m and $\tau_e = 1.0 \times 10^{-10}$ s ($f_e = 1.6$ GHz)²³.

Permittivity dispersion has a pronounced effect on the attenuation rate due, primarily, to a marked increase in effective conductivity in connection with anomalous quadrature permittivity $2\pi f\epsilon''$ near the relaxation frequency f_e . Associated elevation of the attenuation rate is reinforced via the corresponding decrease in effective permittivity, with similar effect on velocity. As illustrated in Figure 33, dispersion occurs largely below the relaxation frequency, with both velocity and attenuation rate becoming nondispersive once again at higher frequencies. Although the relative effect on attenuation rate is clearly more significant for soils having lower real conductivity σ_∞ , Figure 33 indicates that the effect remains appreciable for highly conductive soils.

Nonetheless, the common notion that "high-frequency conductivity" is entirely due to quadrature permittivity is misleading. Consider the relative influence of real conductivity and quadrature permittivity on the effective conductivity $\bar{\sigma} = \sigma' + 2\pi f\epsilon''$. Assuming Debye formulations for σ' (equation 24) and ϵ'' (equation 37), we have $\lim_{f \rightarrow \infty} \sigma'(f) = \sigma_\infty$ (Figure 12b) and $\lim_{f \rightarrow \infty} [2\pi f\epsilon''(f)] = (\epsilon_{dc} - \epsilon_\infty)/\tau_e$ (Figure 16a). As a result, we obtain

$$\chi = \lim_{f \rightarrow \infty} \left[\frac{\sigma'}{2\pi f\epsilon''} \right] = \frac{\sigma_\infty \tau_e}{\epsilon_{dc} - \epsilon_\infty}. \quad (90)$$

In particular, with reference to Figure 33 with $\tau_e = 1 \times 10^{-10}$ s, $\epsilon_{dc} - \epsilon_\infty = 10\epsilon_0$ H/m and $\sigma_\infty = 1, 10, 100$ and 1000 mS/m, we have respectively $\chi = 0.0011, 0.011, 0.11$ and 1.1 , confirming that real conductivity remains a significant contributor to the attenuation rate within the normal range of soil conductivity. Indeed, the relative contribution at a given conductivity level increases substantially as $\epsilon_{dc} - \epsilon_\infty$ decreases with decreasing partial saturation. The relative effect is charted in Figure 34a where $(\epsilon_{dc} - \epsilon_\infty)/\epsilon_0$ is displayed as a function of partial saturation as predicted by the CRIM

²³ A relaxation frequency of 1.6 GHz approximates the the frequency dispersion of partially saturated silts at 25° C as reported by Hoekstra and Delaney (1974).

relation (equation 41) for soils having porosity $\phi = 0.5$ and matrix permittivities $\epsilon'_s/\epsilon_0 = 3.0$ and 7.0 . Dispersion is modelled as due entirely to orientational relaxation of constituent water $\epsilon_{dc} = 80$, $\epsilon_\infty = 5$. Note that χ increases by approximately an order of magnitude as partial saturation decreases from $S_w = 0.5$ to $S_w = 0.1$.

Moreover, because conductivity and permittivity are both directly related to water content, any moisture dependent increase in permittivity will be reflected to some degree via a corresponding rise in conductivity. Figure 34b displays the relation between relative permittivity and conductivity as measured by Smith-Rose (1933), Morey (1977) and Olhoeft (1985). Superimposed, for reference, are empirically derived linear relations with frequency as a parameter (Scott, et al. (1967) and theoretical relations between conductivity, predicted via Archie's relation (equation 19), and permittivity, predicted by CRIM and Topp relations, as functions of partial saturation. Although the precise nature of the relation between conductivity and permittivity is difficult to characterize, particularly for clay soils, the common dependence on partial saturation is evident. Thus, despite conventional wisdom, it follows that electrical conductivity as measured via direct current resistivity or induction electromagnetic methods is a reliable indicator of relative attenuation at radar frequencies, both through its direct influence and as an indirect indicator of moisture dependent permittivity variation.

Compared with the effect of permittivity dispersion, lower frequency conductivity dispersion has a relatively minor influence on the attenuation rate as observed in Figure 33a and a similar effect on velocity (Figure 33b). In fact, conductivity dispersion is typically distributed over a broader bandwidth and potentially of greater magnitude than depicted in Figure 33. We note that the chargeability of soils as summarized in Figure 13 were determined via time domain measurements and inherently underestimate the true chargeability. Seigel (1959) defined chargeability in connection with direct current quadrupole measurements and the transient voltage response arising upon application and subsequent termination of a constant source current as depicted in Figure 35. In particular chargeability is defined as the ratio of voltage between potential electrodes immediately following current termination to that measured immediately prior to termination. Referring to Figure 35a, chargeability is, thus,

$$\eta_0 = \frac{(V_+(\infty) - V_+(0))}{V_+(\infty)} = \frac{V_-(0)}{V_+(\infty)} = \frac{\sigma_\infty - \sigma_0}{\sigma_\infty}, \quad (91)$$

where $V_+(\tau_+)$ and $V_-(\tau_-)$ describe the time dependent potential difference, assuming unit electrode spacing, for charging ($0 < \tau_+ = t - t_+ < t_- - t_+$) and discharging ($\tau_- = t - t_- > 0$), respectively ²⁴. Assuming $t_- = 0$, a generalized time-dependent chargeability follows as

$$\eta(t) = \frac{V_-(t)}{V_{+,0}} = \eta_0 e^{-(\frac{1-\eta_0}{\tau})t}. \quad (92)$$

²⁴ Assuming adequate sampling, Fourier transformation of $V_-(\tau_-)$ yields the associated frequency dependent conductivity spectrum as per equation (22)

In practice, due to inductive effects associated with termination of the source current, chargeability is measured at some fixed time Δt (0.05-10 s) after termination and, consequently, it follows from equation (92) that resulting time domain chargeabilities inherently underestimate η_0 . For example, assuming $\eta_0 = 0.5$ and time constants $\tau = 1.0, 0.1$ and 0.01 , we have $\eta(\Delta t = 1.0\text{ s}) = 0.3, 0.003$ and 1.0×10^{-22} . Clearly, time domain chargeabilities fail to reflect the contribution of higher frequency polarization. It is well established (Bertin and Loeb, 1976; Iliceto, et al., 1982) that observed decay curves are closely approximated by a weighted sum of exponential terms

$$\tilde{\eta}(t) = \sum_{j=1}^m \lambda_j e^{-t/\tau_j}, \quad (93)$$

where, parameters λ_j and τ_j are determined via numerical regression techniques. In practice, a three term approximation is ordinarily sufficient to achieve an acceptable fit. Although Iliceto et al. (1982) demonstrated that exponential decomposition of decay curves yields enhanced discrimination of soil textural classes, as illustrated in Figure 35b, chargeability fields depicted in the same Figure were computed for $\Delta t = 1.28$ s and, thus, as for the data presented in Figure 13, underestimate the actual chargeability $\eta_0 = \eta(0) = \sum_{j=1}^3 \lambda_j$. Frequency domain conductivity spectra reported by numerous authors, including Mitchell and Arulanandan (1968), Olhoeft (1985) and Wensink (1993), indicate that it is not unusual for clays and silts to possess chargeabilities exceeding $\eta_0 = 0.9$.

Finally, the relative influence of magnetic permeability dispersion is illustrated in Figure 36. In addition to conductivity and permittivity dispersion as per Figure 33, superimposed dashed curves include the effect of magnetic permeability dispersion as reported by Olhoeft and Capron (1994) for 20 percent fine iron filings in silica powder. In particular, dispersion is characterized via Cole-Cole parameters $\mu_{dc}/\mu_0 = 1.31$, $\mu_{\infty}/\mu_0 = 1.03$, $\tau_{\mu} = 5.0 \times 10^{-10}$ s and $c = 0.28$. Although the 20 percent iron-silica mixture is substantially more magnetic than ordinary natural soils, we note that associated dispersion of the attenuation rate and velocity is relatively minor compared with permittivity dispersion within approximately the same frequency range. Noting that Figures 31 and 33 were computed under the standard free space assumption $\mu_r = \mu'(f)/\mu_0 = 1$, roughly constant deviation between dashed and solid curves outside the dispersion band provides an indication of the slight, yet potentially significant influence of magnetic permeability variation.

Assuming dispersionless material parameters, analysis summarized in Figures 37-38 indicates that for frequencies substantially lower than the transition frequency f_T , a one percent perturbation in magnetic permeability has precisely the same effect on velocity and attenuation rate as an equivalent perturbation in electrical conductivity. Consistent with previous discussion of Figure 31, the same perturbation in electrical permittivity has comparatively negligible effect at frequencies $f < f_T$. At higher frequencies, magnetic permeability perturbation has a dominant influence on velocity

within the transition region and an effect equivalent to that of electrical permittivity for $f > f_T$, where, consistent with earlier observation, electrical conductivity perturbation has no influence. In contrast, the high frequency attenuation rate is most sensitive to conductivity perturbation while the relatively lesser effect of magnetic permeability perturbation is again equivalent in magnitude to the permittivity effect but of opposite sign.

In general, sensitivity to magnetic permeability dispersion is comparable to that for electrical parameters. In practice, however, natural variation of soil magnetic permeability is typically less than one percent ($\mu_{dc} = 1.00 - 1.01$), depending almost exclusively on ferrimagnetic mineral content and grain size distribution. In comparison, the electrical permittivity of soils spans roughly an order of magnitude ($\epsilon_{dc}/\epsilon_0 = \kappa = 3 - 40$)²⁵, primarily as a function of moisture content, soil texture and temperature. Finally, electrical conductivity of natural soils ranges over several orders of magnitude ($\sigma_{dc} = 0.01 - 1000$ mS/m)²⁶ and depends on a wide range of factors, including moisture content, texture, salinity, cation exchange capacity and temperature. Clearly magnetic permeability has a relatively restricted role in determining the nature of electromagnetic interaction with soils. However, the potential remains and must be evaluated in respect of site specific soil conditions.

4.0 CHARACTERIZATION OF DRES MINE LANE SOILS

As requested a combination of laboratory and field investigations have been conducted to characterize the physical properties of DRES mine lane soils. Laboratory analyses, including textural classification and a combination of dielectric probe and time-domain reflectometry (TDR) measurements were conducted to establish the frequency dependent electrical characteristics of three soil samples as a function of variable moisture content and temperature. Soil sampling was conducted by the Scientific Authority (SA). Two samples of the mine lane aggregate, MLA-1 (start track) and MLA-2 (mid track), and a single sample of the natural prairie soil NS-1 were tested.

Subsequently, a programme of in-situ testing, including TDR, vertical electrical sounding (VES) and frequency domain electromagnetic conductivity measurements, was subsequently conducted at DRES to provide direct connection to laboratory results and assess the degree of soil property variability along the mine lane.

Background environmental data for the DRES-EPG site, as relevant to the interpretation, prediction and assessment of soil electrical property variation are compiled in Appendix II (Bel-MK, 1995).

²⁵ Neglecting low-frequency free charge polarization effects leading to effective permittivities exceeding $\epsilon_{dc}/\epsilon_0 = 1 \times 10^4$.

²⁶ Neglecting high-frequency dielectric polarization effects leading to elevated effective conductivity at frequencies in excess of 1 GHz.

4.1 LABORATORY INVESTIGATIONS

4.1.1 Textural Classification

Textural classification of soils involves the mechanical-hydrodynamic separation and sorting of the bulk soil into constituent soil particle classes, according to established grain-size categories. Classification is based upon the relative proportions of constituent grain-size fractions. A number of standard classification schemes have been devised in connection with engineering and agricultural applications. Recognizing the primary role of silt and clay fractions in controlling the electrical properties of soils, we have developed a specialized textural classification scheme based on the United States Department of Agriculture (USDA) system with emphasis on fine grain soil fractions as summarized in Figure 39.

Standard dry sieve gradations were performed on mine lane aggregate samples MLA-1 and MLA-2. Sieve analysis involves mechanical shaking of the bulk soil sample through a stacked series of standardized wire screen sieves having progressively smaller openings. The mass sample fraction retained on a given sieve is recorded as a function of the standardized mesh gauge. In addition to sieving, a hydrometer analysis was performed on natural soil sample NS-1. The hydrometer method is based on the progressive settling of finer soil particles from suspension in accordance with Stoke's equation for a freely falling sphere within an viscous fluid. Consequently the effective particle size for fine-grained fractions is quoted as the diameter of the equivalent sphere.

Resulting grain size distributions for DRES soil samples are presented in Figure 40. Corresponding textural classifications, according to the recommended classification scheme are as follows:

MLA-1:	well graded	loamy sand and gravel
MLA-2:	well graded	sand and gravel
NS-1:	well graded	loam

4.1.2 Dielectric Probe

To characterize the frequency dependent electrical properties of DRES soils and the dependence upon moisture content, broad-band complex permittivity measurements were acquired using an HP85070A dielectric probe, configured with an HP8752C vector network analyzer, having a bandwidth of 300kHz-3GHz (Hewlett-Packard, 1990). A National Instruments IEEE-488 GPIB card provided interface to HP85070B control software on an IBM pc platform. The apparatus is pictured in Figure 41.

The dielectric probe is, in effect, a terminated or open ended coaxial transmission line (Figure 42). Measurements are acquired by bringing the probe face into firm contact with the flat surface of a solid or semi-solid, or by immersion into a liquid.

Because the radiation resistance of a terminated line is effectively infinite, no appreciable field is radiated. Measurement is achieved via fringe field interaction with the sample under test and, consequently, the sample volume need only be large enough to contain the fringing field ²⁷. At each discrete frequency within a specified measurement band, the network analyzer applies a continuous monochromatic source signal to the dielectric probe. Corresponding fringe field interaction with the sample under test is quantified by measuring the magnitude and phase of the incident and reflected signals. The so-called S-parameter S_{11} , given by the vector ratio of reflected and incident signals, represents the effective complex reflection coefficient, from which the control software computes the effective complex relative permittivity of the sample

$$\vec{\epsilon}_r = \vec{\epsilon} / \epsilon_0 = \frac{\vec{\epsilon}'}{\epsilon_0} + i \frac{\vec{\epsilon}''}{\epsilon_0}, \quad (94)$$

where

$$\vec{\epsilon}' = \bar{\epsilon} = \epsilon' + \frac{\sigma''}{2\pi f} \quad (95)$$

$$\vec{\epsilon}'' = \frac{\bar{\sigma}}{2\pi f} = \epsilon'' + \frac{\sigma'}{2\pi f}. \quad (96)$$

Note the relation between equations (95-96) and equations (62-63). The source frequency is automatically incremented (i.e. swept frequency) and the measurement process repeated to yield a frequency dependent effective complex relative permittivity spectrum.

For single port reflection-based measurements, the HP8752C network analyzer is corrected for systematic measurement errors via a three-standard calibration procedure. Measurements are acquired over the selected bandwidth (200 MHz - 3 GHz) for an open probe, shorted probe and deionized water. Given known values for the corresponding S-parameters, control software computes and stores associated correction factors that are automatically applied to subsequent measurements to remove the effect of systematic instrumental error (Rytting, 1987).

To ensure that the dielectric probe system was properly configured, analyses were conducted on polar liquid standards, including ethanol, methanol, isopropanol and deionized water, over a range of temperatures. Resulting data presented in Figures 43-46 were determined to be consistent with published data (Blackham, 1995; NBS, 1958). In general, we note that measured spectra display dispersion characteristic of polar liquids as described by Debye (equation 35) and Cole-Cole (equation 38) relations.

²⁷ The manufacturer recommends a minimum sample thickness of $z = 20\kappa^{-1/2}$ mm, where $\kappa = \epsilon_{dc}/\epsilon_0$ denotes the approximate dielectric constant of the sample under test Hewlett-Packard, 1993.

To characterize the dependence of complex permittivity on moisture content, soil samples were prepared in a semi-saturated (soil paste) state. Dielectric probe analyses were subsequently performed at regular intervals during progressive evaporation of soil water. Final measurements were acquired following oven treatment of the samples. Procedure is as follows:

- 1 . Determine the mass of M_0 of 100 mm x 15 mm petri dish ²⁸.
- 2 . Sieve samples for approximately 150 ml residual sample passing #10 (2.0 mm) mesh ²⁹
3. Semi-saturated sample via fine mist with thorough mixing (NOTE: avoid free water).
4. Transfer semi-saturated sample (soil paste) to petri dish.
5. Determine combined mass M of semi-saturated sample (soil paste) and petri dish.
6. Perform dielectric probe analysis.
7. Repeat 5-6 daily, or as required, with progressive evaporation of soil water until M reaches a plateau ("room-dry") value.
8. Oven treat sample to incrementally liberate residual water until M reaches a secondary plateau ("oven-dry").
9. Repeat 5-6, as required, with incremental oven treatment.

Dielectric probe analysis (step 6) is conducted according to the following procedure:

1. Clean probe face using lab wipe and deionized water.
2. Conduct single port calibration of the HP85070A probe, employing open, short and deionized water standards.
3. Bring sample surface in firm contact with probe face via adjustable lab jack (NOTE: no quantitative measure of contact pressure was employed).

²⁸ Petri dishes were employed to maximize the ratio of open surface area to volume. Preferential moisture loss from the sample surface (as a fraction of total water mass) remains a significant source of error in measurement of gravimetric moisture content. 60 mm x 15 mm petri dishes may be preferable.

²⁹ Given limited probe contact area, dielectric probe measurements were restricted to the fine grain soil fraction. Manufacturer recommends grain size restriction $\phi \leq 0.3$ mm, noting that repeatability for granular media depends on density variation (Hewlett-Packard, 1993). Our experiments confirm that inclusion of larger grains can substantially skew the measured permittivity from the corresponding mean. However, mean spectra were observed to remain stable with inclusion of larger grains, confirming that bulk sample permittivity is dominated by the fine grained constituents. It is our opinion that $\phi < 2.0$ mm (#10 mesh) represents a practical tradeoff between instrumental precision and accurate characterization of bulk soil electrical permittivity.

4. Initiate swept-frequency measurement cycle.
5. Repeat steps 3 and 4, repositioning the probe on the sample surface, for a total of five replications ³⁰.

Average frequency dependent (200 MHz-3 GHz) complex relative permittivity spectra are displayed in Figures 47, 48 and 49 for DRES soils MLA-1, MLA-2 and NS-1, respectively, with gravimetric moisture content w as a parameter. Specific moisture dependence at 1 GHz, 2 GHz and 3 GHz is illustrated in Figures 50, 51 and 52. In each case, the experimental data display a well defined relation between complex permittivity and gravimetric moisture content $w = M_w/M_s$, where M_w and M_s denote mass of water and mass of dry solid fractions, respectively. In addition, the observed dependence is reasonably stable over the frequency range 1-3 GHz. "Best-fit" (minimum square error) exponential functions $a_w e^{b_w w}$ have been computed to facilitate subsequent prediction of complex effective permittivity on the basis of in-situ soil moisture measurements on DRES soils. In general, exponential fits closely predict measured data. Second and third order polynomials gave inferior results. It is anticipated that a generalized exponential $a_w c_w^w$ could yield a minor reduction in misfit.

4.1.3 Time Domain Reflectometry

To facilitate a direct connection between laboratory analyses of DRES soils and in-situ characterization, and recognizing that dielectric probe instrumentation is not well suited for field application, the laboratory analysis program was augmented to include parallel time domain reflectometry (TDR) measurements utilizing a Tektronix 1502C TDR and special purpose test probe-cell as depicted in Figures 53 and 54. TDR measurements yield a broad-band estimate of the dielectric constant (real relative permittivity) by determining the effective propagation velocity of a guided wave within the sample under test (Topp et al, 1980). The Tek1502C generates a RF step pulse ³¹ which is guided through air along a coaxial test lead and subsequently through the sample medium along the parallel probe set. Like a radiated radar pulse, the TDR pulse is partially reflected upon encountering contrasts in electromagnetic impedance. In particular, the pulse is reflected at the connection between the test lead and the probe, at the interface between air and sample and, ultimately, at the probe's termination. The TEK1502C displays the composite returned waveform as a function of calibrated time via a built in storage oscilloscope. The waveform can also be internally digitized and down-loaded via an RS232 port to a portable computer for subsequent

³⁰ Replicate probe analyses were acquired to permit subsequent statistical reduction of random measurement error associated with sample heterogeneity, surface moisture variability and non-uniform contact pressure.

³¹ Manufacturer specified pulse rise time $\Delta t_r < 200$ ps (Tektronix, 1994).

analysis³². Given known probe length ℓ within the sample, and by measuring elapsed time between the arrival of reflections from the air/sample interface t_0 and the probe termination t_e , the effective electromagnetic propagation velocity is obtained as

$$v_{TDR} = \frac{2\ell}{t_e - t_0}. \quad (97)$$

Assuming that the soil under test is effectively dielectric $\tan \delta \ll 1$ over the TDR bandwidth³³, an estimate of the effective dielectric constant follows as

$$\kappa_{TDR} = \frac{v_0}{v_{TDR}} = \left[\frac{v_0 (t_e - t_0)}{2\ell} \right]^2, \quad (98)$$

where $v_0 = 1/\sqrt{\mu_0 \epsilon_0}$ denotes electromagnetic wave velocity in free space.

TDR analysis procedure is identical to that set out above for dielectric probe measurements, with the exception that the standard sample vessel for TDR measurements is a 600 ml Pyrex beaker. The parallel probe set comprises $d = 2.0$ mm diameter stainless steel probes, having length $\ell = 8.5 - 9.5$ cm and fixed separation $s = 3.0$ cm. Probes are connected to the TEK1502C via a short parallel lead and, subsequently, via a balun and coaxial cable. The associated intrinsic probe impedance (Ohms) follows from a general relation for parallel transmission lines (Krause, 1984; Gray, 1972)

$$Z = \frac{120}{\sqrt{\kappa}} \ln \left(\frac{2s}{d} \right), \quad (99)$$

where $\kappa = \epsilon_{dc}/\epsilon_0$ denotes the effective dielectric constant of the background medium (eg. air/soil). A recent paper by Suwansawat and Benson (1999) provides a useful systematic analysis of the influence of various TDR test cell design parameters on measured dielectric constant.

As anticipated, comparison of TDR measurements on sieved and unsieved samples indicates (in contrast with dielectric probe measurements) that repeatability of TDR measurements is relatively unaffected by removal of coarse sample fractions. To facilitate probe insertion and to avoid potential damage to pyrex cell, samples were sieved to remove coarse gravel retained at $\geq 1/2$ inch mesh.

Prior to testing DRES soils, TDR analyses were performed on polar liquid samples in connection with calibration of the dielectric probe as described in the previous section. Acquired waveforms displayed in Figure 55 provide a qualitative indication

³² Acquisition control, data transfer and subsequent analysis were facilitated via WATTDR software, developed by the Waterloo Centre for Groundwater Research (Redman, 1995).

³³ Figure 54.5 displays the TDR step pulse and corresponding amplitude spectrum as characterized by the manufacturer. The specified maximum frequency is $f_{max} = 0.35/t_r \approx 1.75$ GHz, where $t_r = 200$ ps denotes the corresponding rise-time of the step pulse. We note that inclusion of dominantly low frequencies calls to question the theoretical basis for a dielectric assumption²²

of temperature dependence. A range of quantitative analysis techniques were investigated for automated identification of reflection onsets. In addition to standard linear regression based approaches, a variety of minimax and inflection point techniques were investigated. None of these automated techniques were determined to be satisfactorily robust. Rather, TDR waveform analyses reported here were conducted manually. In particular, on the basis of reference traces presented in Figure 51, a constant start (air-sample interface) time $t_0 = 23.20$ ns was established for the test cell, with termination (end) time t_e , picked manually for each waveform as the first up-going and sustained break from minimum amplitude.

Figure 56, displays resulting temperature dependent dielectric constant estimates for ethanol, methanol, isopropanol and distilled water. It was anticipated that correlation of TDR permittivity-temperature relations with corresponding dielectric spectra in Figures 43-46 would identify a consistent effective TDR measurement frequency. However, despite appropriate magnitudes for associated permittivities and well defined relations between dielectric constant and temperature, the effective measurement frequency does not appear to be consistent. For instance, decreasing permittivity with increasing temperature observed for methanol implies an effective measurement frequency below 1 GHz, while the relatively low dielectric constants measured for deionized water suggest an effective frequency exceeding 1 GHz. In short, findings suggest that the effective TDR measurement frequency is significantly influenced by the electrical properties of the sample under test.

Moisture dependent TDR waveforms acquired on DRES soil samples are displayed in Figure 57. As per dielectric probe analyses, all TDR measurements were acquired under temperature controlled lab conditions at 19-22° C. As previously described reference time $t_0 = 23.2$ ns was assumed constant for the test cell, with end time t_e , picked manually for each waveform as the first up-going and sustained break from minimum amplitude. Corresponding dielectric constant estimates are displayed in Figure 58 as functions of gravimetric soil moisture content w . Corresponding dielectric probe data and associated optimum-fit exponential relations are overlaid for comparison. Although there is, in general, very good agreement between the two methods. TDR estimated dielectric constant is systematically higher and displays a marginally lower rate of increase with gravitational moisture content. Optimized exponential expressions $\epsilon' = a_w e^{b \cdot w}$ characterizing TDR permittivity-moisture relations permit a quantitative comparison with corresponding dielectric probe data.

Taken together and in consideration of data misfit, best fit permittivity-moisture relations as determined by TDR and dielectric probe techniques for sample MLA-1 and MLA-2 constitute gross bounds on expected variability of mine lane electrical permittivity, both spatially and as a function of soil moisture. In addition, we note that field-based TDR estimates of broad-band dielectric constant for DRES soil can be translated, via intercalibrated permittivity-moisture relations to obtain equivalent complex effective permittivity parameters as established by dielectric probe analyses

on representative samples MLA-1 and MLA-2. In particular, on computing mean coefficients $\bar{a}_w$ and $\bar{b}_w$ for optimum dielectric probe (DP) and TDR permittivity-moisture relations for mine lane samples MLA-1 and MLA-2, we obtain the respective reference relations

$$\bar{\epsilon}_{DP}(w) = 1.5864 e^{0.1398 w}, \quad (100)$$

$$\bar{\epsilon}_{DP}'(w) = 0.0579 e^{0.1708 w} \quad (101)$$

and

$$\bar{\epsilon}_{TDR}(w) = 2.7561 e^{0.1127 w}. \quad (102)$$

Now, given a TDR field-measured relative effective permittivity $\bar{\epsilon}_{TDR}$, we solve the previous equation for the associated laboratory calibrated moisture content

$$w_{TDR} = 8.8731 \ln(0.3628 \bar{\epsilon}_{TDR}) \quad (103)$$

and, subsequently, the corresponding complex relative permittivity via substitution in equations (100-101).

In connection with the use of foregoing results to predict the moisture-dependent electrical permittivity of DRES soils, it is emphasized that both TDR and dielectric probe results presented above for DRES soils were acquired at an approximately constant temperature of 19-20° C. As per previous discussion in Section 2.2 (Figure 22), the electrical permittivity of soils depends significantly on temperature. Consequently, the temperature range over which foregoing analysis applies is restricted. To assess the nature of temperature dependence, TDR analyses were conducted on samples MLA-1 and NS-1 to characterize the influence of temperature on the effective dielectric constant. In particular, samples were frozen at a temperature of -10° C, with the TDR probe set and thermometer in place. TDR waveforms were subsequently acquired as bulk sample temperature gradually equilibrated with ambient laboratory conditions. Gravimetric moisture levels were $w = 13.30\%$ and $w = 26.35\%$ for samples MLA-1 and NS-1, respectively. Resulting data are presented in Figure 59. In general, the temperature dependence for DRES soils is consistent with that observed in Figure 22. Measured dielectric constants are approximately twice corresponding "dry" values at -10° C with crystalline water having relatively little influence on the soil's effective polarizability. Dielectric constant increases only gradually with increasing temperature below 0° C and, subsequently, rises abruptly as water changes state in proximity of 0° C. Dielectric constant stabilizes once again at higher temperatures, reaching a plateau value of approximately $\kappa = 16$ above 10° C for MLA-1. The rate of transition is lower for sample NS-1, approaching a plateau value of approximately $\kappa = 11$ for temperatures above about 15° C³⁴. It follows that permittivity-moisture relations

³⁴ A lower transition rate for NS-1 is attributed to enhanced liquid water retention for $T < 0^\circ \text{C}$ and adsorption-related viscosity for $T < 0^\circ \text{C}$ in association with finer texture and correspondingly higher specific surface.

based on analyses at 20° C are restricted to temperatures $T > 10^\circ \text{C}$ and $T > 15^\circ \text{C}$ for samples MLA-1 and NS-1, respectively. Additional temperature dependent analyses are required to generalize the dependence of permittivity-temperature relations on moisture content at lower temperatures.

Beyond characterization of DRES soils, foregoing results demonstrate that the TDR method can provide reliable measurement and monitoring of soil dielectric constant for in-situ prediction-assessment of GPR performance and real-time calibration of land mine detection and recognition software. Having said this, a major drawback of the standard TDR method, compared with the dielectric probe technique is the lack of full spectral characterization of the complex effective permittivity. As previously indicated, the standard TDR technique yields a broadband estimate of dielectric constant, rather than a complete frequency-dependent spectrum³⁵. Related time-domain methods have been developed for parallel estimation of the electrical conductivity on the basis of direct analysis of waveform attenuation (Nadler et al., 1991; Heimovaara, 1992; Baker and Spaans, 1993). The approach is illustrated in Figure 60a, depicting TDR waveforms acquired in saline solutions having progressively higher salinity and proportionally higher electrical conductivity as per equation (18). Relative attenuation is quantified via the relation $\alpha_{TDR} = V_0/V_{LT}$, where V_0 and V_{LT} denote initial and "late-time" TDR voltage levels as indicated in Figure 60b. TDR estimated effective electrical conductivity σ_{TDR} follows as

$$\sigma_{TDR} = \frac{1}{120\pi\ell} \frac{Z_0}{Z_u} (2\alpha_{TDR} - 1), \quad (104)$$

where Z_0 and Z_u denote the characteristic impedance of the probe set (equation 99) and the output impedance of the TDR transceiver, respectively (Topp et al, 1988; Ferrè et al, 1998). Although this approach has not been pursued due to restricted project scope, relatively little effort would be required to derive associated electrical conductivity estimates from existing TDR waveforms.

We note, however, that as for standard TDR estimates of the dielectric constant, the foregoing method for determining effective electrical conductivity yields a broad-band estimate, as opposed to full spectral characterization. Consequently, acknowledging the evident potential of the TDR method for in-situ characterization of high-frequency soil electrical properties, we recommend that future research be directed toward the development of TDR waveform inversion techniques for estimation of frequency dependent complex permittivity spectra. In fact, such methods were reported in the physical chemistry literature more than two decades ago (Giese and Tiemann, 1975)

Breifly, the TDR waveform $w(t)$ is represented as the time convolution of the TDR pulse $h(t)$ with a time series $s(t)$ representing the time-variable system response

³⁵ Historically, TDR techniques developed in connection with soil science were utilized primarily for calibrated moisture content and salinity assessment without requirement for spectral characterization (Topp et al., 1980).

or reflectivity as a function of characteristic impedance encountered by the TDR pulse during propagation along the transmission line. The mathematical description is

$$w(t) = h(t) * s(t) = \int_{-\infty}^{+\infty} h(t - \tau) s(\tau) d\tau, \quad (105)$$

where * denotes time convolution and τ is an integration variable effecting relative time shift between $h(t)$ and $s(t)$. The corresponding frequency domain relation is

$$W(f) = H(f)S(f), \quad (106)$$

where the TDR waveform and its frequency spectrum are related via the Fourier transform

$$W(f) = \int_{-\infty}^{+\infty} w(t) e^{-i2\pi ft} dt, \quad (107)$$

with similar realtions

$$H(f) = \int_{-\infty}^{+\infty} h(t) e^{-i2\pi ft} dt \quad (108)$$

$$S(f) = \int_{-\infty}^{+\infty} s(t) e^{-i2\pi ft} dt. \quad (109)$$

for the pulse spectrum $H(f)$ and the system transfer function $S(f)$. In principle, the complex electrical permittivity is obtained by Fourier transforming the acquired TDR waveform and subsequently dividing by the known pulse spectrum, yielding the system transfer function or S-parameter

$$S(f) = \frac{W(f)}{H(f)}, \quad (110)$$

which is in turn a function of the complex effective electrical permittivity $\tilde{\epsilon}^*(f)$ ³⁶. The precise form of $S[\tilde{\epsilon}^*(f)]$ depends on probe configuration and associated impedance match at the transmission lead. $\tilde{\epsilon}^*(f)$ is subsequently obtained via a range of inversion or parameter optimization techniques based on forward modelling of the S-parameter as a function of Debye or Cole-Cole permittivity spectra.

Recently, foregoing techniques have been applied to characterize the frequency dependent complex effective permittivity of soils with promising results (Heimovaara, 1994; Heimovaara et al., 1996; Friel and Or, 1999). Although reported experimental analyses have been conducted exclusively with coaxial probes, we expect that similar results can be obtained for data acquired with the parallel probe configuration employed here. In fact, it is anticipated that utilization of the parallel probe configuration

³⁶ In practice, computational complexities arise in connection with signal windowing, discrete Fourier transformation of the TDR step pulse $h(t)$ and spectral division where $H(f) \rightarrow 0$.

in combination with the appropriate S-parameter relation may lead to more robust complex permittivity estimates than obtained by using an ideal coaxial S-parameter model to approximate data acquired with multi-prong radial probe configurations as described by Heimovaara (1994) and Friel and Or (1999).

4.1.3 Electrical Permittivity Models

To assess the use of physical-empirical models for predicting electrical permittivity-moisture relations, experimental results described in the previous section were compared with three common mixing models. In particular, Figures 61-62 display electrical permittivity as a function of volumetric moisture ψ_w as predicted by CRIM (equation 41), a pure volumetric mixing formula, the Wang and Schlugge (WS) model (equations 44-47), a combined physical-empirical formulation and the purely empirical relation due to Topp *et al.* Corresponding data for DRES soils are displayed as functions of volumetric moisture as estimated via the formula

$$\psi_w = w\rho_b, \quad (107)$$

where w denotes gravimetric moisture and $\rho_b = M_s/V_T$ represents dry bulk density, defined as the ratio of the mass of dry soil solids M_s to total soil volume V_T , including air, water and solid fraction. In particular, volumetric moisture for soils in Figures 61-62 is based on indicated bulk densities measured for oven dry samples.

It is preferable to characterize soil electrical permittivity as a function of volumetric moisture to remove the influence of grain density which has relatively little or no bearing on electrical characteristics. Rather, as previously emphasized it is the amount (dipole moments per unit volume) and related distribution of water within the pore volume (adsorbed, capillary, free) that ultimately determines the soil's electrical polarizability. In general, lower bulk densities for fine textured soils reflect a correspondingly higher porosity (Figures 3 and 63). Note that the relative disparity between gravimetric moistures for sample NS-1 (loam) and samples MLA-1, MLA-2 (sands) is much reduced in terms of volumetric moisture.

Returning to Figures 61-62, there is clearly general agreement between measured and predicted relative permittivity. Although none of the models provides a fully adequate representation of the measured data, there is reasonably good correspondence at low moistures, particularly with the CRIM relation. Real relative permittivity as measured via the dielectric probe is, in all cases, substantially higher than predicted at volumetric moistures exceeding approximately $\psi_w = 25\%$. Although quadrature effective permittivity of sand samples (MLA-1, MLA-2) is similarly underestimated at higher volumetric moistures, there is significantly better agreement with the Wang and Schlugge model for the loam sample NS-1. In fact, while Figures 61-62 display permittivity-moisture relations predicted by the WS model using default parameter values for ψ_s^* (equation 46) and Γ , limited experimentation with these adjustable parameters indicates that the WS model is sufficiently flexible to fit the measured data within a reasonable tolerance.

In general, limited investigation of selected physical-empirical models confirms the potential of these relations for predicting soil electrical characteristics on the basis of more readily available physical properties, including texture, moisture, CEC and other standard soil survey data. In fact, it is our opinion that existing models can be utilized in conjunction with empirically derived connecting relations to predict gross soil electrical properties on the basis of existing soil classifications and prevailing environmental conditions.

4.2 FIELD INVESTIGATIONS

To provide connection to laboratory characterization of DRES soil properties, a programme of geophysical field investigations was undertaken at DRES to characterize in-situ soil electrical properties. In situ measurements included TDR, vertical electrical resistivity soundings (VES) and induction electromagnetic conductivity. Measurement locations were selected by the Scientific Authority and distributed on the basis of ongoing field testing of the ADM-ELTA radar, to investigate a suspected connection between ELTA system performance and variable soil electrical characteristics. Fieldwork was carried out during the period 26-28 August 1998. Weather conditions were clear with a daily maximum temperature of approximately 20-25° C. Fieldwork commenced daily at 08:00 MDT and ended approximately 17:00 MDT. The following is a factual description of field results originally reported to the Scientific Authority by Fax dated 10 October, 1998.

4.2.1 Time Domain Reflectometry (TDR)

TDR measurements of dielectric constant were made at nine sites along the center-line of the mine lane (measurement locations are referenced to a temporary site datum). Depth dependent dielectric constant was investigated at each site by acquiring TDR waveforms using 10 cm, 20 cm and 30 cm probe sets. For each case, the measured dielectric constant represents a depth averaged value over the length of the corresponding probe set. Results are presented in Figure 64.

In general, we note that measured dielectric constant $\kappa = 5 - 7$ is consistent with laboratory results (Figure 58, MLA-1, MLA-2) at a gravitational moisture level of approximately $w = 5 - 10\%$. Relatively minor depth variation is observed except near each end of the track where 30 cm readings indicate a significantly higher dielectric constant, presumably in connection with increased soil moisture levels at depth.

4.2.3 Electrical Resistivity (VES)

To investigate the potential influence of free-charge electrical conductivity, coincident Wenner quadrapole resistivity soundings were acquired in conjunction with TDR measurements. Resistivity soundings were acquired using a Nilsson M-400 resistance meter. The unit generates a 97 Hz square wave voltage signal that is applied to the outer electrodes (current dipole) of the Wenner array as indicated in Figure 65.

The M-400 detects the resulting potential difference between the inner electrodes (potential dipole). The effective soil resistance is determined by balancing a calibrated internal bridge circuit. Measured soil resistance R is corrected for Wenner electrode geometry to obtain the apparent or average resistivity $\rho_a = 2\pi aR$, where a denotes the constant electrode interval. To a first approximation, the effective sounding depth is approximately $1.5a$. Consequently, as the electrode spacing increases, the measured apparent resistivity reflects the influence of progressively deeper soils.

Measured apparent resistivity profiles are presented in Figure 66 for electrode intervals of $a = .10, .20, .30, .40, .50, .70$ and 1.0 m. Resulting data have subsequently been computationally inverted (Zohdy, 1989) to yield an optimized depth dependent resistivity structure for each sounding. Inversion results are presented in Figure 67, overlayed on a depth-dependent colour-contour cross-section through the centerline of the DRES mine lane. Actual soundings are displayed as shaded depth-profiles. Note that computer generation of the associated depth-section involves inverse-distance-weighted interpolation between measurement locations and subsequent fitting of a minimum curvature surface to the interpolated grid. Consequently, given the relatively sparse nature of measured data (i.e. distance between adjacent VES sites), depicted depth-dependent electrical resistivity is not well constrained and discretionary interpretation is advised. In fact, as per following discussion, continuous electromagnetic conductivity data indicate abrupt lateral variation in subsurface electrical conductivity and contradict interpolated resistivity data at several locations along the DRES mine lane.

With regard to potential correlation between spatial variability in measured galvanic resistivities and ELTA radar performance, we note that the effective electrical conductivity ³⁷. $\bar{\sigma} = \sigma' + 2\pi f\epsilon''$ (equation 62) comprises both the free-charge electrical conductivity σ' measured via the galvanic resistivity method, and a bound-charge conductivity $2\pi f\epsilon''$ associated principally with high-frequency orientational polarization of constituent water. Although bound-charge conductivity does not contribute significantly to the maintenance of a sustained charge current density at low frequencies employed for galvanic resistivity measurements, the contribution of free-charge conductivity can have a significant influence at radar frequencies as previously described in Section 3.0. In addition, because conductivity and permittivity are both directly related to soil moisture content, moisture dependent variation of the electrical permittivity will be reflected to some degree via corresponding electrical conductivity. As a result, electrical conductivity as measured via galvanic resistivity or induction

³⁷ Note that although electrical resistivity $\rho = \rho' + i\rho''$ is the inverse of electrical conductivity $\sigma = \sigma' + i\sigma''$, both parameters are complex valued. Consequently due to significant dispersion at low frequencies due to free-charge polarization processes (Section 2.1), real electrical conductivity computed as the reciprocal of electrical resistivity may not be precisely equivalent to measured resistivity

electromagnetic methods can be a practical indicator of relative attenuation and velocity variability at radar frequencies.

4.2.3 Electromagnetic Conductivity

In addition to localized TDR and VES measurements, the electromagnetic (EM) induction method was employed to provide semi-continuous mapping of electrical conductivity variations along the mine lane. In particular, high density EM conductivity measurements were acquired along three mine lane profiles, one along the track center-line, another 1.5 m south of center-line and a third 1.5 m north of center-line. The sample interval along profile was 0.5 m.

Basic principles of the electromagnetic induction method are illustrated in Figure 68. An alternating current is fed through a wire transmitter coil, producing a time-varying magnetic field B_p that penetrates the ground and induces secondary electrical currents J_e within subsurface materials in accordance with Faraday's law of electromagnetic induction. These so-called eddy currents, in turn, give rise to a secondary magnetic field B_s that is sensed together with the primary field by the receiver coil. Induced current flow and, consequently, the strength of the secondary magnetic field are approximately proportional to subsurface electrical conductivity³⁸. Measurements were acquired using a Geonics EM-38 in both vertical and horizontal dipole modes.³⁹ Depth sensitivity (and range) is significantly different for the two modes as illustrated in Figure 69. In horizontal mode, the EM-38 has maximum sensitivity at surface. In vertical mode, the EM-38 is relatively insensitive to surface soil conductivity and possesses maximum sensitivity at approximately 0.35 m. Nominal penetrations (70% cumulative response) for horizontal and vertical dipole operation are .75 m and 1.5 m, respectively.

Corresponding electrical conductivity scans acquired along the DRES mine lane are displayed in Figure 70. Again, we note that contour level and colour shading are approximate and involve interpolation and subsequent fitting of a minimum curvature polynomial surface. However, due to the relatively high density EM-38 coverage along transects, the resulting image is well constrained. Consequently, where relative EM conductivity in Figure 70 contradicts estimates based on VES data in Figure 67, indicated EM conductivity is to take precedence.

³⁸ Strictly, the linear relation between secondary magnetic field intensity and electrical conductivity holds only for materials having an induction number $B = s_c/\delta_s \ll 1$, where s_c (=1.0 m for EM-38) denotes the coil separation and $\delta_s = (\pi f \mu \sigma)^{-1/2}$ ($\delta_s = 1/\alpha$ with $\alpha \approx (\pi f \mu \sigma)^{1/2}$ for a conductive medium as per Section 3.0) represents the effective skin depth. Here, f (=14.6 kHz for EM-38) denotes the transmitter frequency. In short, the intercoil separation must be small in comparison with the skin depth. We note that this condition is almost invariably valid for natural soils.

³⁹ Instrument is operated with transmitter and receiver coils oriented such that axes are vertical or horizontal relative to ground surface.

Like the galvanic resistivity method, the electromagnetic induction technique is based on the flow of conduction currents within the subsurface and, consequently, yields a measurement of the free-charge electrical conductivity σ' as opposed to bound-charge electrical conductivity $2\pi f\epsilon''$ associated with high-frequency polarization effects. Again, analysis presented in Section 3.0 is consistent with our experience that electromagnetic conductivities are a reliable predictor of GPR performance. In particular, experience with GPR in connection with the detection, monitoring and remediation of groundwater contamination, confirms that free-charge electrical conductivity has a significant influence on the progressive attenuation and dispersion of the radar signal. Consequently, we are not surprised that EM conductivity scans of the DRES mine lane show significant correlation with ELTA radar performance, as indicated by the Scientific Authority.

5.0 MINE SURROGATE GPR SIGNATURES

To investigate the effect of climate-driven soil property variation on the GPR signature of landmines, an experimental test bed has been constructed for acquisition of GPR scans over standard DSTO landmine surrogates. Repeat scans acquired under variable environmental (moisture-temperature) conditions yield a qualitative assessment of related signature variation.

Two separate phases of experiments (VTS-1 - VTS-2) were undertaken. The initial investigation (VTS-1) consisted of preliminary scans to assess the background response of the test bed followed by two sets of replicate GPR scans at 500 MHz, 900 MHz and 1 GHz. Test-bed background scans and target deployment were carried out on 17 December, 1997, with the initial round of target scans (VTS-1-1) acquired on 18 December, 1997. A second round of scans (VTS-1-2) was conducted on 28 January, 1998. Due to unusually high precipitation levels during the period November-January, 1997-98 and project schedule constraints, both rounds of GPR testing were conducted under saturated soil conditions, yielding an insufficient data set for comparative analysis of environmental conditions. Consequently, results were augmented via a second phase of testing (VTS-2) under relatively dry soil conditions during the period September-November, 1998. The second phase comprised three rounds of replicate GPR scans at 900 MHz and 1 GHz⁴⁰. Phase 2 fieldwork was conducted on 2 September, 1998 (VTS-2-1), 18 September, 1998 (VTS-2-2) and 3 November, 1998 (VTS-2-3). In connection with each round of GPR testing, in-situ TDR and EM-38/VES data were acquired to characterize soil dielectric constant and electrical conductivity. In addition, soil samples were acquired for laboratory determination of gravimetric moisture.

⁴⁰ Plans to extend testing to 2 GHz were not realized as development and testing of the new antenna system had not been completed by the manufacturer.

5.1 Vancouver Test Site (VTS)

The test bed was established during November, 1997 on a property adjacent to Golder Associate's Burnaby B.C. office at 4260 Still Creek Drive. A rectangular void, measuring approximately 3.0 m (east-west) $\times$ 10.0 m (north-south) $\times$ 1.5 m (depth) ⁴¹, was excavated into local till soil (Figure 71) and subsequently backfilled with a uniform well graded sand fill. Laboratory-measured textural and electrical properties of the fill are summarized in Figures 72 and 73. In general, electrical characteristics are similar to those for the DRES mine lane aggregate (MLA-1, MLA-2).

5.2 Mine-Like Targets - Characteristics and Deployment

Reference targets for test-bed calibration experiments include three DSTO standard antipersonnel landmine surrogates: ST-AP(1) (M14); ST-AP(2) (PMN); ST-AP(2) (PMN-2), an anti-tank mine surrogate (ATM), a granite cobble and a metallic calibration target. Reference targets are pictured in Figure 74, with corresponding schematic diagrams for antipersonnel mine surrogates depicted in Figure 75a.

Corresponding dielectric probe data acquired on surrogate surfaces is summarized in Figures 76 and 77. In particular, outstanding electrical permittivity characteristics are observed, apparently in connection with an approximately 2.0 mm layer of rubber overlying PVC/ABS, on the upper surface of PMN and PMN-2 surrogates. Bakelite, an internal component of the PMN-2, and the ATM surface are characterized by intermediate permittivities, with yet lower values associated with PVC/ABS surfaces. Bulk paraffin (internal PMN-2) and parowaxTM possess the lowest measured permittivities.

Figure 75b depicts the spatial deployment of reference targets for phases VTS-1 and VTS-2. Spacing and alignment were adjusted to ensure a consistent transect alignment and reduce interference between target signature. As indicated, all scans were conducted for a burial depth of 10 cm to target interface. In order to isolate the influence of environmental variables and to minimize "hole effects", targets, once deployed, were left in place throughout the duration of associated testing.

5.3 Data Acquisition - Instrumentation and Methodology

GPR scans were acquired in continuous mode, utilizing a GSSI SIR-2 digital GPR system (GSSI, 1994). Transducers were in contact with ground surface and moved manually at approximately constant rate along the transect with fiducial marks recorded at 0.5 m intervals to provide constraints for subsequent spatial correction. Quoted antenna frequencies are free-space centre frequencies. Corresponding centre frequencies within subsurface fill material are lower, depending on coupling characteristics

⁴¹ A 1.5 m test-bed depth was designed to avoid interference with wide angle target signatures at low frequencies.

as a function of frequency-dependent electrical properties. During the initial round of testing for each phase, optimum gain settings were identified and stored. Identical acquisition parameters were employed for subsequent rounds to isolate the effects of environmentally controlled electrical property variation. If necessary, supplemental scans were acquired with adjusted gain settings to compensate for soil variations. Data were recorded in digital format for subsequent reduction, processing and analysis.

In-situ TDR measurements were acquired using a Tektronix 1502C, with 10, 20 and 30 cm probe sets to quantify depth-dependent soil dielectric constant. Corresponding low-frequency soil electrical conductivity was measured via a Geonics EM-38 (VTS-1) or via vertical electrical resistivity soundings VES (Nilson M-400) with $a=10$, 20 and 30 cm (VTS-2).

5.4 Environmental Conditions

Figures 78 and 79 display mean temperature and precipitation data for periods December, 1997-January, 1998 (VTS-1) and September, 1998-November, 1998 (VTS-2) as recorded by the Greater Vancouver Regional District, Air Quality Department at Confederation Park Monitoring Station approximately 2 kilometers north-east of the test site and at similar elevation.

5.5 Results and Interpretation

As expected results obtained during the initial phase VTS-1-1 / VTS-1-2, under generally saturated soil conditions display relatively little variation as illustrated by comparison of 1 GHz scans in Figure 80. Target signatures are slightly "tighter" in connection with a minor increase in dielectric constant as indicated in Figure 81. A more evident signature for the M14 in the initial scan VTS-1-1 is attributed to hole-effect rather than surrogate signature variation. Slightly lower conductivity for VTS-1-1 (potentially attributable to progressive leaching of soluble) yields only minor effect on inductive coupling. Finally, although the influence of temperature variation is difficult to decouple from soil moisture related effects, we note that increased temperature generally reinforces the effect of higher moisture content.

A more interesting finding in connection with phase VTS-1, is the observation of a strong resonance signature for the PMN-2 surrogate at 500 MHz. Although PMN and PMN-2 surrogates produce relatively similar GPR signatures at higher frequencies, as displayed in Figure 82 for 1 GHz, the PMN-2 exhibits an exceptionally high amplitude resonance response at 500 MHz compared with a relatively weak and transient response for the PMN as illustrated in Figure 83. We are aware of recent research on this topic (Chen, et al., 1995; Chen and Peters, 1997) and are persuaded by results in Figure 83 of the significant potential of resonance signatures for recognition and discrimination of landmines. In view of this potential, it is our opinion that investigation of broad-band swept frequency radar acquisition techniques is an appropriate research direction and could lead to development of viable new

methods for landmine recognition (Crice, 1994; Noon *et al.*, 1994; Moriyama *et al.*, 1998).

In contrast with results of VTS-1 investigations, acquired under generally saturated soil conditions, VTS-2 trials were conducted under relatively dry conditions and span a significant range of soil moisture variation. In particular, with reference to Figure 79, we note that VTS-2-1 scans were acquired under practically "dry conditions", VTS-2-2 scans followed the first significant autumn precipitation event, and VTS-2-3 scans were acquired after significant accumulated precipitation during early October and an ensuing dry period during the following two weeks. All VTS-2 trials were completed prior to the onset of winter rainfall during early November.

Figures 84-94 display corresponding representative signatures for each reference target as acquired at 900 MHz and 1 Ghz. Average soil electrical properties for the respective suite of scans are indicated below the associated signature. In contrast with nearly consistent signatures obtained under saturated soil conditions (VTS-1), pronounced variation in target signature is observed in response to associated soil property variation at low moisture levels. In particular, signatures are relatively low amplitude under "dry" conditions (VTS-2-1) compared with signatures recorded at higher moisture levels (VTS-2-2, VTS-2-3). An associated decrease in effective propagation velocity, with increasing dielectric constant, also gives rise to a general increase in the apex curvature of target signatures, as previously discussed in connection with Figure 81.

Perhaps more obvious is the significant variation in inductive coupling characteristics observed for VTS-2-2 scans in connection with substantially higher electrical conductivity. High-amplitude banding associated with enhanced coupling is a general source of interference and effectively masks weaker signatures. Noting that VTS-2-2 scans were acquired following the first significant autumn precipitation, increased conductivity may be attributable to the re-mobilization of salts precipitated in connection with evapotranspiration processes at the fill surface during preceding summer months. This ascertainment is supported by an observed return to lower conductivities at higher moisture content, during VTS-2-3 investigations, suggesting that residual salts had been subsequently leached and anomalous ion concentrations diluted via substantial intervening precipitation.

Again the relative influence of temperature variation is difficult to assess. Gradually decreasing temperatures during the period of VTS-2 investigations, should theoretically lead to lower effective electrical conductivity and dielectric constant, offsetting the opposite trend associated with increasing soil moisture. It follows that moisture effects dominate the competing influence of temperature variation.

As proposed, the foregoing analysis of test-bed GPR signatures, and associated dependence on environmentally controlled soil electrical property variation, has been restricted to qualitative analysis of unprocessed data. However, we have concurrently experimented with a number of quantitative analysis techniques, including complex

signal attributes and polarization analysis. On the basis of limited investigation of these methods, it is our opinion that combined computer-based quantitative signal characterization techniques can be effective for landmine recognition and discrimination and deserve further investigation and development.

6.0 SUMMARY AND RECOMENDATIONS

The original request for proposals focused principally on the development of standard methods and procedures for combined field and laboratory-based characterization of frequency-dependent soil electrical properties. More specifically, emphasis was placed on the application of standard methods to characterize the physical-electrical properties of DRES mine lane soils as a function of frequency and moisture content.

These objectives have been addressed via a combination of laboratory and in-situ testing as described in Section 4.0. In addition to requested dielectric probe analyses, a complete parallel suite of laboratory analyses have been conducted to assess and calibrate time-domain reflectometry (TDR) techniques for subsequent field application. Although findings clearly demonstrate the potential of TDR methods for accurate in-situ measurement and monitoring of soil electrical properties, limitations are also identified. In particular, while TDR apparatus is relatively inexpensive, robust and field portable, standard waveform analysis techniques are restrictive, yielding broad-band effective electrical properties, as opposed to full spectral characterization of the complex effective permittivity. As previously discussed with the Scientific Authority, we recommend that future research and development activities be directed toward the investigation of TDR waveform inversion techniques for automated estimation of frequency dependent complex permittivity spectra. With improved analysis software, the TDR method could be an indispensable tool for in-situ prediction and assessment of GPR performance potential and for real-time input to landmine detection and recognition software. Moreover, we believe that similar information can be derived via related analysis of surface reflectivity data.

In the meantime, combined dielectric probe and TDR analyses of DRES soils, according to established procedures reported here, has yielded a range of empirically derived relations for prediction of moisture-dependent complex relative permittivity. Field-based TDR estimates of broad-band dielectric constant for DRES mine lane soils can be translated, via intercalibrated permittivity-moisture relations presented in section 4.1.3, to obtain equivalent complex effective permittivity parameters as established by dielectric probe analyses on representative samples MLA-1 and MLA-2. As indicated, however, these relations are only valid for a restricted temperature range. Extension to lower temperatures, where the polarizability of constituent water varies substantially, would require additional laboratory work to establish and calibrate the dependence of permittivity-moisture relations on temperature. Rather, recognizing the inability to precisely simulate field conditions in the laboratory, it is

our opinion that resources would be more efficaciously directed toward development of fully effective in-situ techniques as previously recommended.

A secondary aim of the current research programme, as set out in the original request for proposals, was experimental investigation of the influence of environmentally-driven soil property variations on the GPR signatures of mine-like targets. In order to isolate environmental effects and to facilitate efficient long term monitoring of target signatures, a full scale test-bed was constructed locally as described in Sections 5.0-5.1. Following laboratory characterization of the uniform sand fill, six reference targets, including four standard landmine surrogates were deployed, with associated signatures subsequently monitored in connection with variable environmental conditions. A total of five repeat investigations were conducted in two phases with soil conditions ranging from dry (9.1 %/vol) to saturated (31.8 %/vol). Continuous radar scans were acquired at frequencies of 500 MHz, 900 MHz and 1 GHz, with resulting target signatures digitally recorded for subsequent analysis.

In general, qualitative analysis described in Section 5.5 confirmed strong correlation between target signature and environmentally-driven soil electrical property variation. Three principal effects were observed:

1. Nonlinear signature amplitude increase with increasing soil moisture, with relative effect diminishing at higher soil moisture levels.
2. Increased apex curvature with elevated soil moisture in connection with increased effective permittivity and associated velocity decrease.
3. Increased inductive coupling-modulating effects in association with increased effective conductivity.

Results emphasize that landmine recognition and discrimination software must be real-time adaptive for time and space-dependent soil electrical property variation. Assuming that real-time-estimated complex effective permittivity can be input via discrete TDR or continuous surface reflectivity inversion, appropriate adaptation of mine signature templates requires robust quantitative models for prediction of electrical property-dependent mine signatures. Moreover, optimum signature attributes need to be identified and investigated. Although findings presented in Section 5.0 are limited to a qualitative descriptive analysis, concurrent experimentation with a variety of quantitative analysis techniques indicates that complex signal attributes, polarization characteristics and AVO (amplitude versus offset) properties are potentially valuable signature characterization attributes and deserve further investigation and development.

Finally, though not explicitly stated, the original request for proposals implies the need for a general review of soil phenomenology and, particularly, soil electromagnetic properties in connection with potential development of criteria for predicting GPR performance potential on the basis of traditional soil data base information. Moreover,

recognizing that a rigorous and comprehensive theoretical description of soil electromagnetic properties is an essential foundation for meaningful analysis, modelling and interpretation of experimental data, the author has undertaken an extensive review of relevant scientific literature, spanning soil science, geophysics and physical chemistry. Surprisingly, it is discovered that description of bulk electromagnetic properties of soils is largely absent in standard treatments of soil scientific principles. Similarly, discussion of soil electromagnetic properties is relatively rare in standard geophysical references, where the emphasis is largely on the physical properties of rocks. Moreover, where soil electrical or magnetic properties are addressed in the geophysical literature, there is a tendency to neglect related soil properties and processes. In response, Section 2.0, represents an attempt to unite existing theoretical descriptions of soil electromagnetic parameters, related soil physical properties and processes and relevant reported data. Although beyond the scope of required work, we believe that this review provides a vital framework for interpreting reported experimental findings.

Ultimately, we are confident that the comprehensive description of soil electromagnetic properties presented in Sections 2.0 and 3.0 could provide a foundation for prediction and forecasting of GPR performance potential on the basis of standard soil classification schemes and other existing sources of soil survey data. In fact, for certain regions of the developed world, a variety soil electromagnetic property data is directly available as per Figure 95. Presumably there is similar, perhaps classified, information for for large parts of the globe. In addition, there is ever increasing capability to gather related data via airborne and satellite remote sensing (Wagner *et al.*, 1999; Oevelen and Hoekman, 1999). Where soil electromagnetic properties are known, corresponding GPR performance potential could be evaluated in accordance with a range of quantitative criteria. For instance, Figure 96a displays a map of GPR potential for soils interpretation as compiled by Doolittle (1987) on the basis of effective ground conductivities depicted in Figure 95. Here relative performance criteria are simply electrical conductivity thresholds. Similarly, where sufficient soil survey data, including texture, bulk density, moisture and minearalogical composition are available, theoretical models could be employed to predict associated electromagnetic properties (eg. Wang and Schmugge, 1980; Dobson *et al.*, 1985), with corresponding performance criteria subsequently evaluated.

Where soil physical property data are not directly available, prediction of GPR performance is more complicated and subject to greater uncertainty. However, recent research reported by Collins (1992) and Doolittle and Collins (1995) concludes that soil maps and related classification systems alone are a sufficient basis for assessing GPR performance. As indication of this potential, note the close relation between soil taxonomy in Figure 96b and corresponding electrical conductivity in Figure 95b. In general, soil classification schemes are closely related to soil formation factors, including climate and vegetation as indicated by significant correlation between soil taxonomy and corresponding distributions of mean annual precipitation and natural

vegetation depicted in Figure 97. General relations between climate, vegetation and common soil classifications are charted in Figure 98. Additional soil formation factors include topography, parent material and soil maturity (Jenny, 1941; Birkeland, 1974). In combination, these factors lead to systematic differentiation of related soil properties, including texture, moisture, clay content, porosity, salinity, CEC, organic content, pH and Eh. As described in Section 2.0 of this report, it is precisely these factors that control and determine soil electromagnetic characteristics.

Consequently, although no explicit relations exist between soil classes and electromagnetic properties, it is our opinion that generally applicable empirical relations can be established between soil classes and representative electromagnetic property ranges. In view of potential benefit in connection with the development and fielding of appropriate mine detection technologies, we recommend that DND, in cooperation with TTCP partners, consider undertaking research and development of a global soil-climate data base and related geographic information system (GIS) for prediction of soil electromagnetic properties. We envision that an appropriate research and development programme would comprise four separate phases:

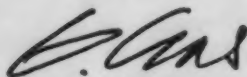
1. Data base development and management: Location, compilation and organization of raw data, including (a) existing soil electromagnetic property data (electrical conductivity, dielectric constant, magnetic susceptibility); (b) related soil property data (texture, moisture, clay content, salinity, CEC, density, organic content, pH and Eh, etc.); (c) existing soil classifications and (d) existing global climate-meteorological data.
2. Data integration and standardization: Standardization of physical, chemical and meteorological data units to SI (Système International d'Unités) and translation-standardization of local-regional soil taxonomy-classifications.
3. Research, development and standardization of theoretical-empirical models relating soil property-climate data and associated electromagnetic properties. Empirical derivation of templates relating soil classes and associated soil properties for subsequent prediction of corresponding electromagnetic properties via theoretical-empirical models.
- 4.0 Development and Calibration of GIS: initially for management of standardized global soil data base and ultimately for prediction of space-time-dependent soil electromagnetic properties. Significant training, calibration and field truthing of the GIS would be required.

It is anticipated that the foregoing programme would involve substantial international cooperation between industry, government and academic research sectors, and would require approximately 3-5 years from initial data-base development to operational GIS.

We trust that the foregoing report satisfies your current requirements and would be pleased to discuss recommendations for further work at your convenience. Should you require additional information or clarification, regarding reported findings or recommendations, please contact the undersigned.

Yours truly,

GOLDER ASSOCIATES LTD.

A handwritten signature in dark ink, appearing to read 'G. Cross', written in a cursive style.

Guy Cross, Ph.D.
Senior Geophysicist

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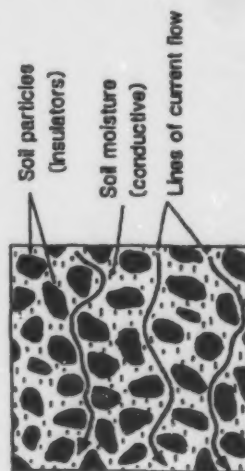
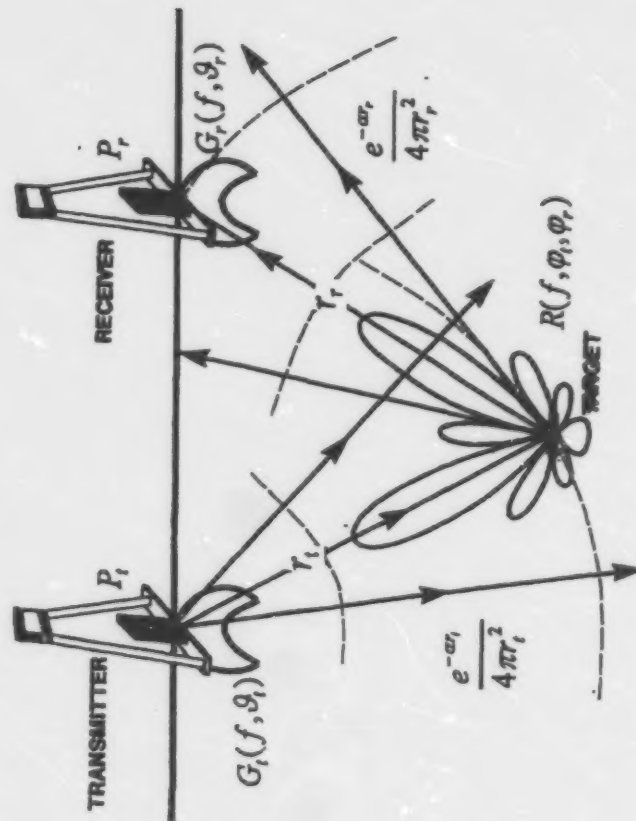
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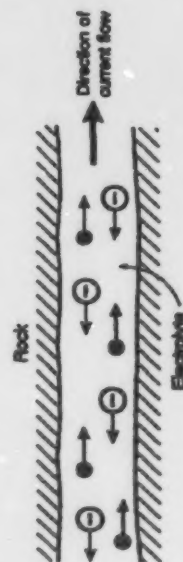
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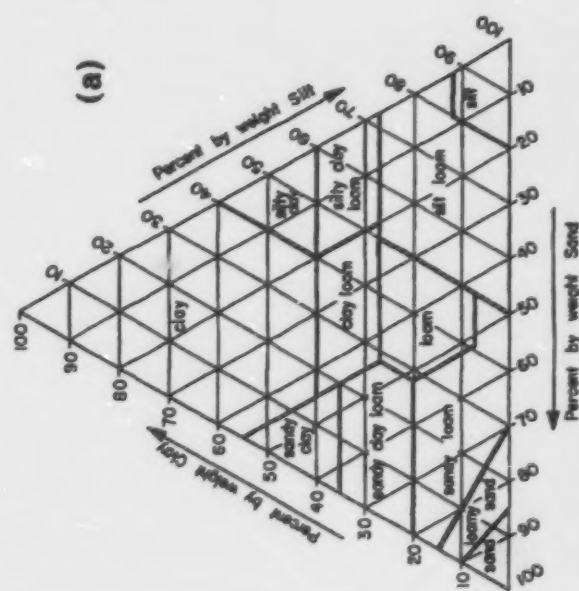


Schematic of current flow in soil sample.

(Ward, 1990)



(Reynolds, 1997)

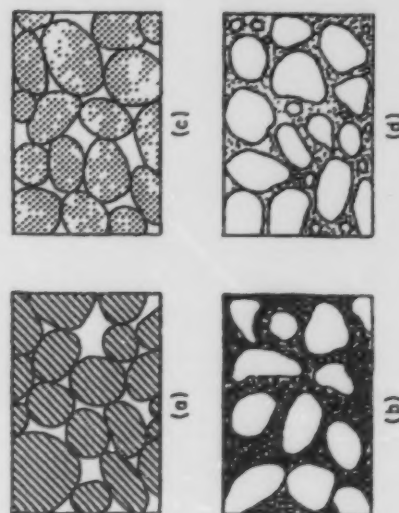
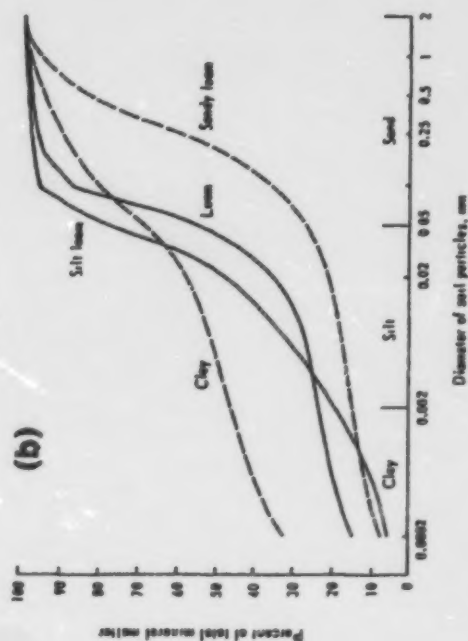


Textural triangle, showing the percentages of clay (below 0.002 mm), silt (0.002-0.05 mm), and sand (0.05-2.0 mm) in the basic soil textural classes.

United States Department of Agriculture	CLAY	SILT	SAND				GRAVEL	
			Very fine	Fine	Med	Coarse	Very coarse	
	0.002		0.05	0.10	0.25	0.5	1.0	2.0 mm

Representative porosity ranges for sedimentary materials (after D. K. Todd [12])

Material	Porosity, %
Soils	50-60
Clay	45-55
Silt	40-50
Medium to coarse mixed sand	35-40
Uniform sand	30-40
Fine to medium mixed sand	30-35
Gravel	30-40
Gravel and sand	20-35



(d)

Relation between texture and porosity. (a) Well-sorted sedimentary deposit having high porosity; (b) poorly sorted sedimentary deposit having low porosity; (c) well-sorted sedimentary deposit consisting of pebbles that are themselves porous, so that the deposit as a whole has a very high porosity; (d) well-sorted sedimentary deposit whose porosity has been diminished by the deposition of mineral matter in the interstices

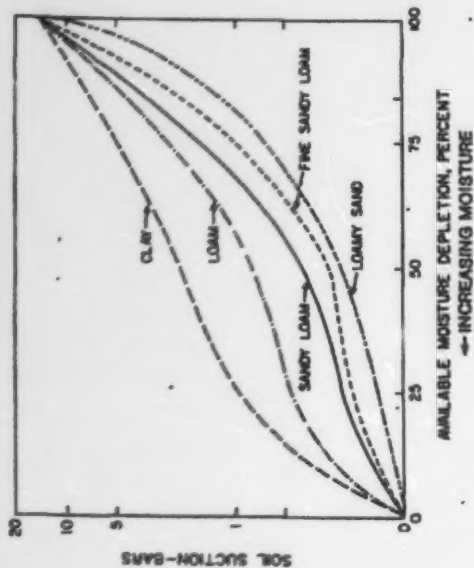
(Freeze and Cherry, 1979)

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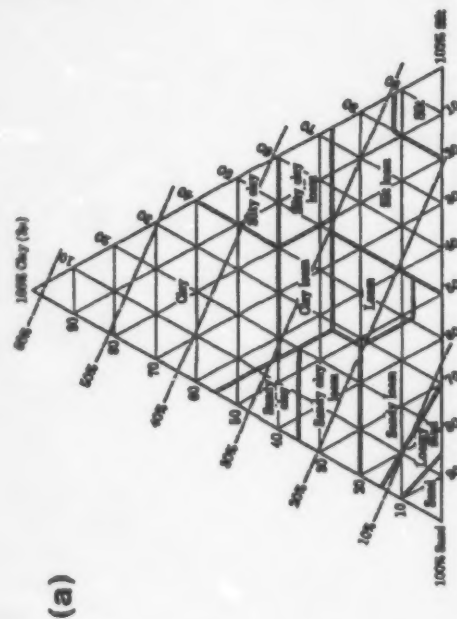
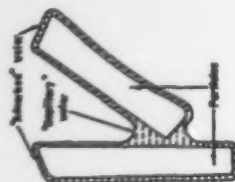


SOIL POROSITY - TEXTURE

(Richards and Richards, 1957)

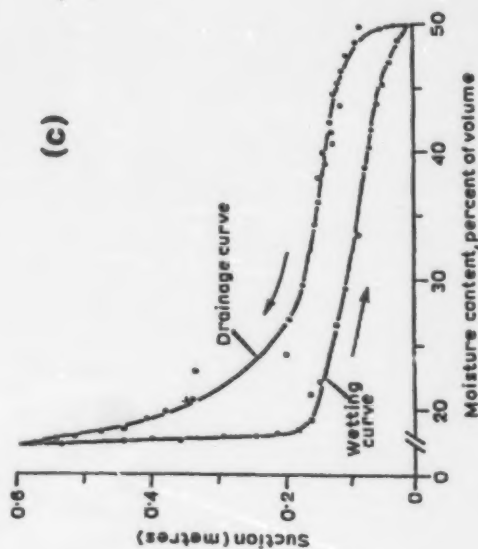


(b)



(a)

Soil textural classes plotted on a triangular diagram.²² The dashed lines are values for moisture equivalent calculated from the equation, moisture equivalent = 0.028 sand + 0.25 silt + 0.61 clay. (Taken from Bodman and Mahmud,⁷ Fig. 3, © 1932, The Williams & Wilkins Co., Baltimore, and Jenny and Raychaudhuri,¹⁶ Fig. 5.) The moisture equivalent approximates field capacity; thus the plot gives a general relationship between soil-texture classes and field capacity for soils low in organic matter. (Birkeland, 1974)



(c)

(Pitly, 1979)

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SOIL MOISTURE RETENTION

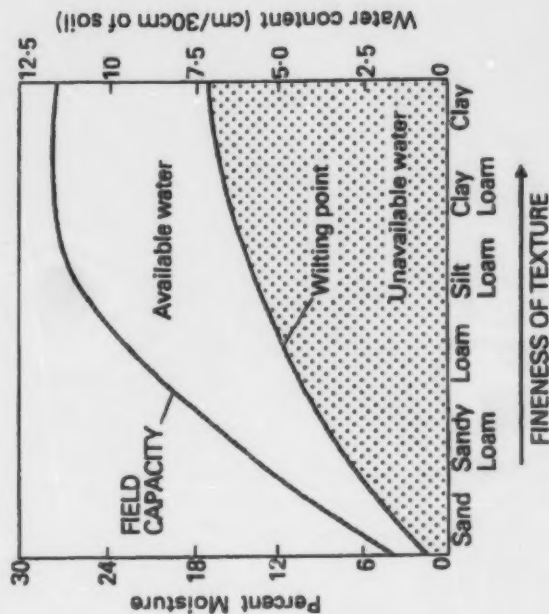
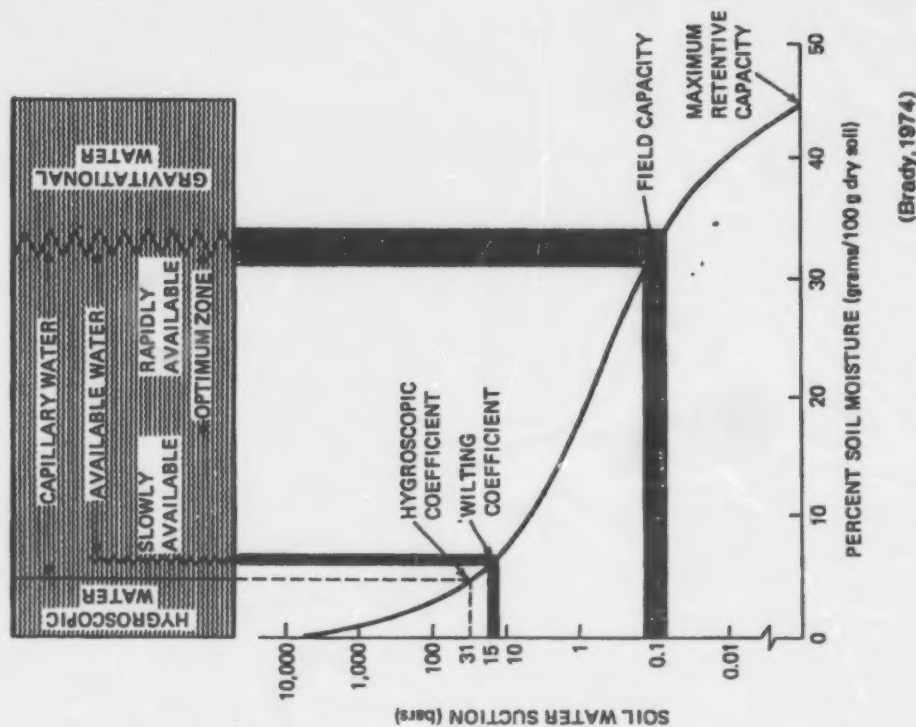
Figure

4

(Hansenbullaier, 1978)

Relationship between soil texture and estimated available-water content of soils as judged from differences in the soil water content at matric potentials of $-1/3$ rd and -15 bars. (Data from various sources).

Textural class	Water content at $\frac{1}{3}$ of available water		Estimated available water
	%	-15 bars	
Silt loam	32	10	%
Silt loam	22	4	22
Loam	29	8	21
Silty clay	36	17	19
Silt loam	28	11	17
Sandy loam	20	4	16
Clay	40	25	15
Silty clay	37	23	14
Loam	21	9	12
Clay	42	31	11
Clay loam	22	12	10
Sandy loam	15	6	9
Sandy loam	14	6	8
Sandy loam	11	5	6
Clay loam	25	20	5
Sandy loam	7	3	4



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SOIL MOISTURE - TEXTURE

Chemical Constituents of Ground Water

Major Constituents Minor Constituents

Common Cations	Equivalent Weight	Common Anions	Equivalent Weight
Calcium (Ca)	20.04	Carbonate (CO ₃)	30.00
Magnesium (Mg)	12.16	Bicarbonate (HCO ₃)	61.01
Sodium (Na)	23.00	Sulfate (SO ₄)	48.03
Potassium (K)	39.10	Chloride (Cl)	35.46
		Nitrate (NO ₃)	62.01
		Fluoride (F)	
		Selenium (Se)	
		Iron (Fe)	
		Aluminum (Al)	
		Silica (SiO ₂)	
		Boron (B)	

From: Todd (1959)

$$\sigma = 96500[C_1M_1 + C_2M_2 + \dots] = 96500 \sum C_iM_i$$

where C_i = no. of gram equivalent weights of i^{th} ion per 10^6 cm^3 of water

M_i = mobility of i^{th} ion in meters per second per volt per meter.

Mobility of common ions at 25°C (after Keller and Fricknecht)

(b)

ion	Mobility (m ² /sec V)
H ⁺	36.2×10^{-8}
OH ⁻	20.5×10^{-8}
SO ₄ ⁻	8.3×10^{-8}
Na ⁺	5.2×10^{-8}
Cl ⁻	7.9×10^{-8}
K ⁺	7.6×10^{-8}
NO ₃ ⁻	7.4×10^{-8}
Li ⁺	4.0×10^{-8}
HCO ₃ ⁻	4.6×10^{-8}

Waters of Various Chemical Types

	1	2	3	4	5	6	7	8	9
	Snow	Well	Well	Well	Well	Well	Well	Acid Hot Spring	Basic Spring
SiO ₂	0.2	135	42	36	7.2	14	14	216	3970
Al	0.06	...	...	0.1	...	...	...	0.02	56
Fe	0.02	0.06	0.03	0.0	0.33	0.0	0.01	33	1.8
Mn	...	...	0.35	0.0	...	...	...	...	0.06
Ca	0.3	5.2	0.4	173	225	686	5.6	185	7.3
Mg	0.3	1.0	0.3	32	41	204	13	32	2.6
Na	0.43	28	51	40	37	462	3.2	6.7	10,900
K	0.1	3.6	3.0	2.6	...	...	5.1	0.5	24
Li	...	...	...	...	...	...	...	...	135
NH ₄	...	...	...	...	...	...	...	...	3.2
HCO ₃	4.0	30	120.0	218	267	72	89	...	1.48
CO ₃	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
OH	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
SO ₄	0.7	7.8	10	248	190	2130	5.0	1570	1490
Cl	0.6	3.5	7.2	135	94	830	3.5	3.5	7180
F	0.0	12	0.0	0.1	0.3	1.5	0.1	1.1	3.0
Br	...	...	...	...	...	...	...	...	11
I	...	...	...	...	...	...	...	...	5.7
NO ₃	0.2	0.3	2.4	7.5	518	...	0.5	0.0	0.0
PO ₄	...	...	...	0.0	...	...	0.0	...	4.3
B	...	0.3	...	...	...	...	...	...	242
Dissolved solids	5.4	309	177	820	1306	4400	94	...	31,200
Condensates	10	164	214	...	...	5610	141	4570	35,000
pH	6.0	7.6	7.9	7.4	...	7.3	7.3	1.9	11.6
Temperature, °F	107	78	...	...	...	73	63	...	54
H ₂ S	...	...	...	...	...	...	...	...	1000

From: Davis and DeWiest (1966)

Chemical Analyses of Selected Ground Waters in California

(After Donsen 19)

Num- ber	EC × 10 ⁴ B, at 25° C	Major Constituents, mg/l						Per Cent Na	Water Chem °
		Ca	Mg	Na	CO ₃ +HCO ₃	Cl	SO ₄		
1	200	1.41	0.44	0.99	1.08	0.24	0.23	23	Good
2	370	0.31	0.06	2.43	1.20	0.06	0.67	92	Unsuitable
3	790	0.34	0.02	7.28	2.39	2.47	2.45	95	Unsuitable
4	900	0.51	0.40	8.81	8.87	1.13	1.03	36	Unsuitable
5	1000	...	1.30	2.00	8.10	1.00	2.60	73	Unsuitable
6	1270	0.25	0.20	3.94	3.46	2.73	4.47	30	Unsuitable
7	1760	0.71	2.14	0.08	13.67	12.04	1.80	92	Unsuitable
8	3550	0.60	11.40	6.70	12.90	2.80	23.00	46	Unsuitable
9	4330	1.63	13.37	16.71	27.39	9.55	41.74	49	Unsuitable

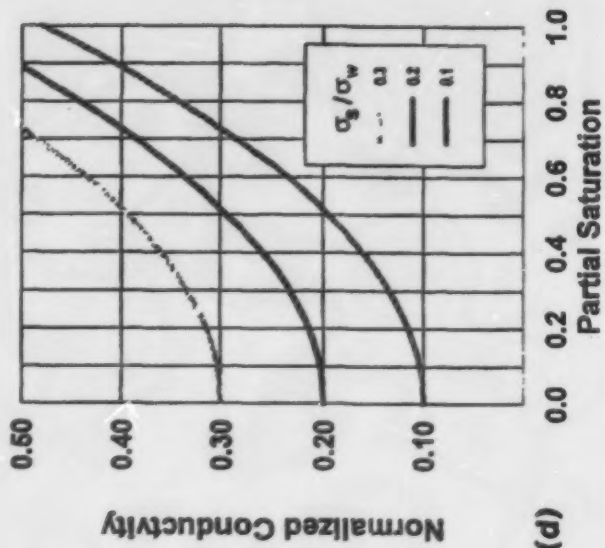
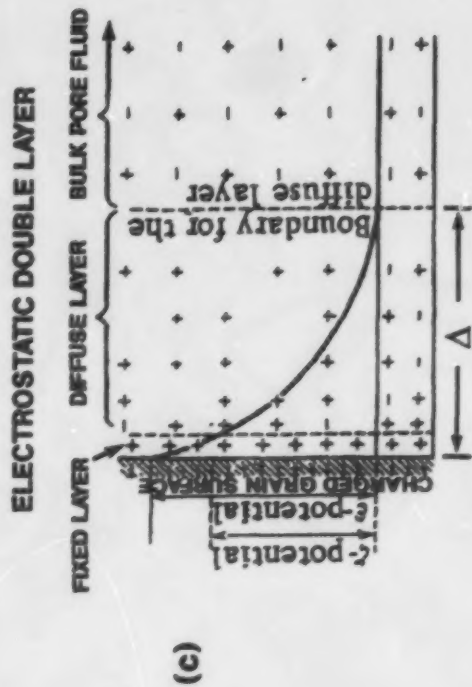
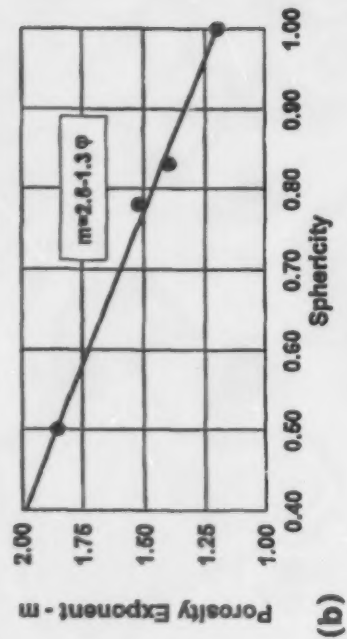
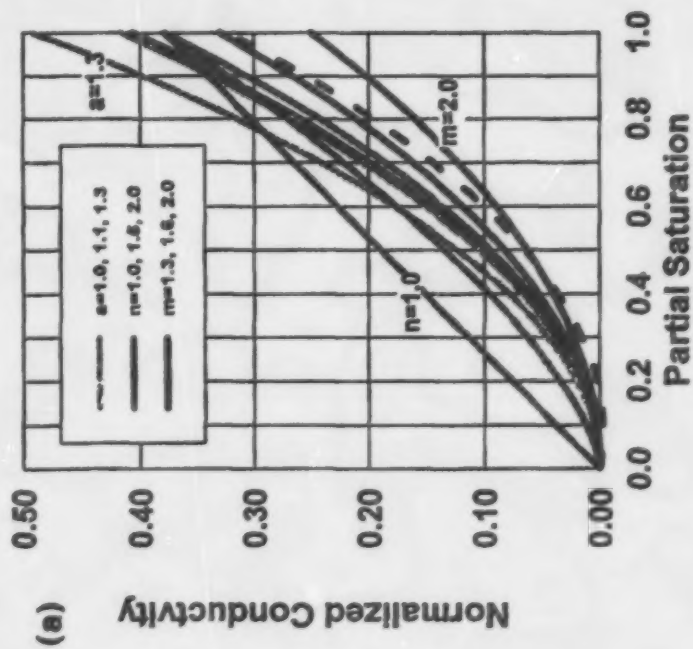
From: Todd (1959)

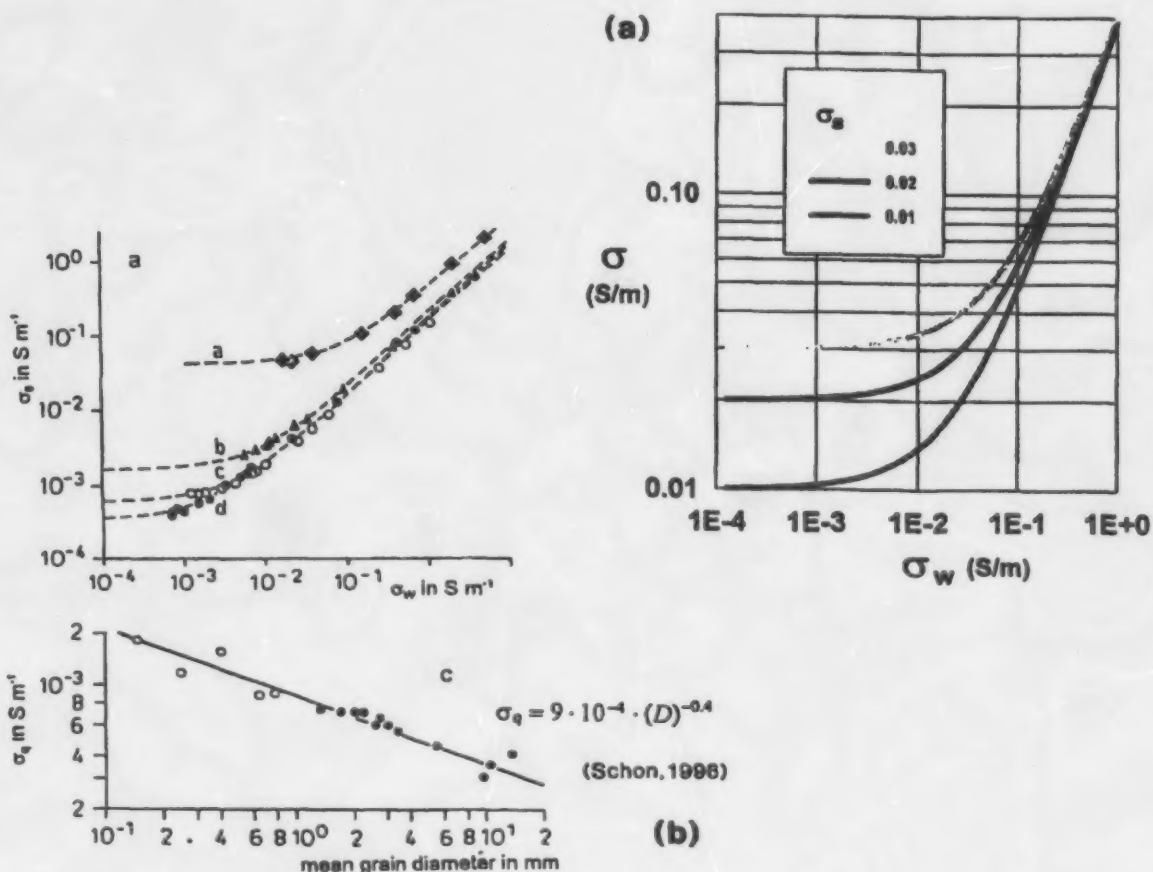
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PORE WATER ELECTRICAL CONDUCTIVITY

Figure



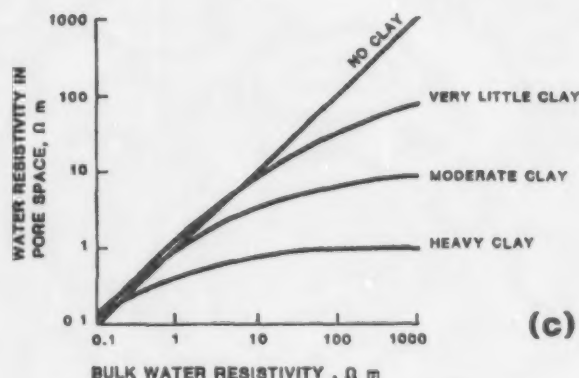


Electrolytic and interface conduction components of unconsolidated sediments (Börner, 1983, Börner and Schön, 1984, Schön, 1984)

a) Examples of the dependence of rock conductivity σ_r upon the brine conductivity σ_w (a - clay, b, c, d - sand, gravel)

b) Formation factor F versus porosity Φ (the three lines are calculated after the structured model with the angle α as parameter)

c) Interface conductivity σ_i versus mean grain diameter D



(Keller, 1987)

(b) The cation exchange capacity (CEC) and colloid content of five soils of different textures (after L. R. Webber (6)).

Soil Texture	Organic Matter %	Clay %	CEC* me/100 g
Sand	1.7	7	6.3
Sandy loam	3.2	13.2	13.7
Loam	4.9	16.8	20.2
Silt loam	5.4	18.4	24.0
Clay loam	5.5	31.2	27.2

*The cation exchange capacity of a soil is expressed in terms of milliequivalents per 100 grams of soil (meq/100 g). A milliequivalent is defined as 1.0 milligram of hydrogen or the amount of any other element that will combine with or displace it. The milliequivalent weight of any element may be found as follows:

$$\frac{\text{atomic weight of element}}{\text{valence of element} \times 1000} = \text{milliequivalent weight}$$

(McNeil, 1980)

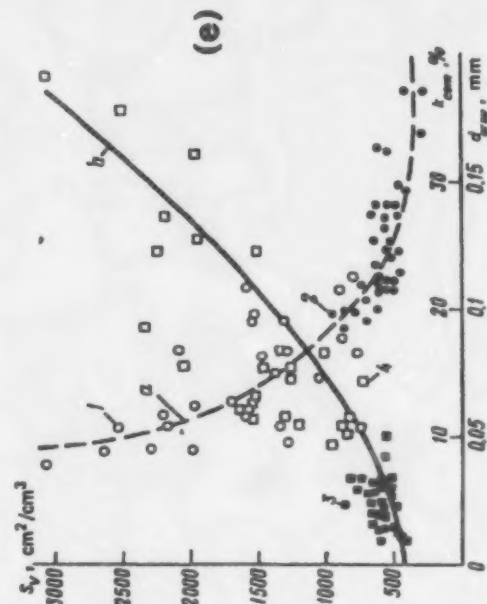
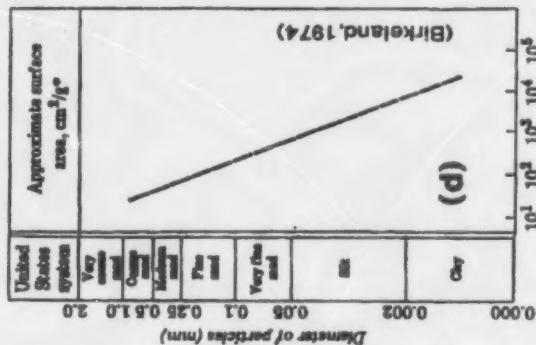
(a) Exchange capacity of common clays (after Keller and Frischknecht (5))

Clay	Exchange Capacity
Kaolinite	3 to 15 m-equiv/100 g
Halloysite . 2H ₂ O	5 to 10
Halloysite . 4H ₂ O	40 to 50
Montmorillonite	80 to 150
Illite	10 to 40
Vermiculite	100 to 150
Chlorite	10 to 40
Attapulgite	20 to 30

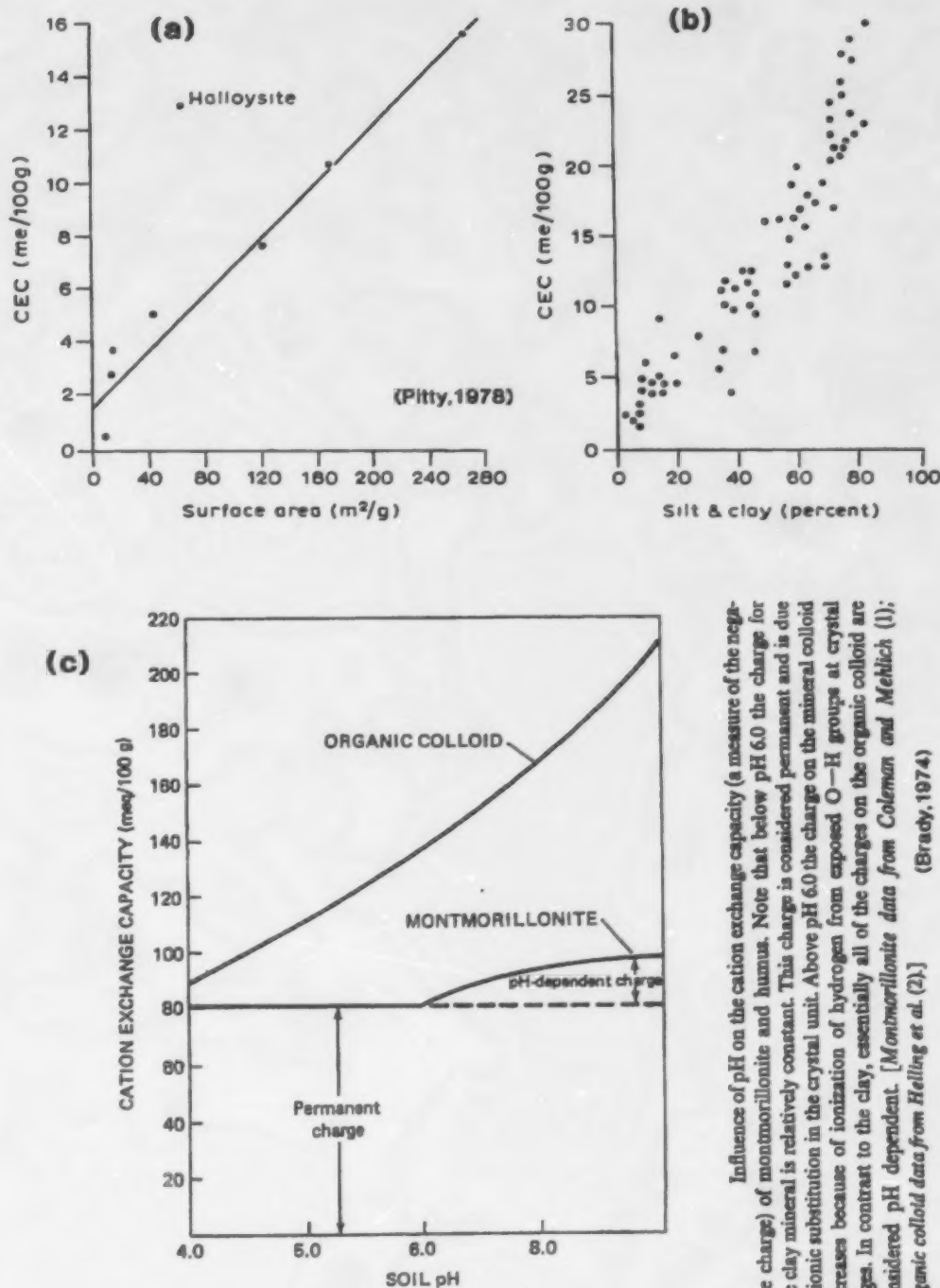
(c) Ranges of specific surface of clays and soils in square meters per gram

Clay, montmorillonite	300-800
Clay, illite	60-120
Clay, kaolinite (soil)	20-40
Clay, kaolinite (vermiculite)	10-20
Clay, soil-clay fraction in Middle West	300-400
Clay, from glacial calcareous till	150-200
Clay soil	170-250
Silty clay loam	120-200
Silt loam	50-150
Loam	50-100
Sandy loam	10-40
Silt soil	5-20

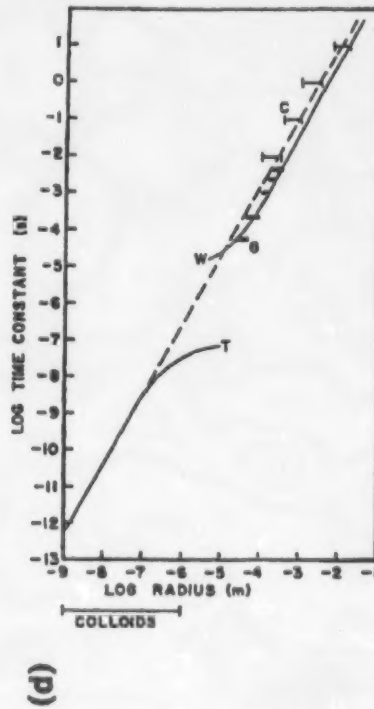
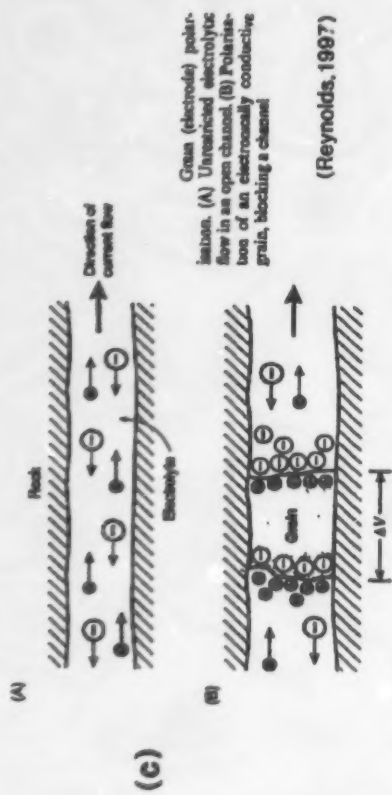
(Kohnke, 1968)



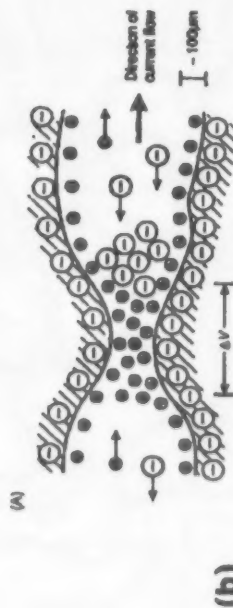
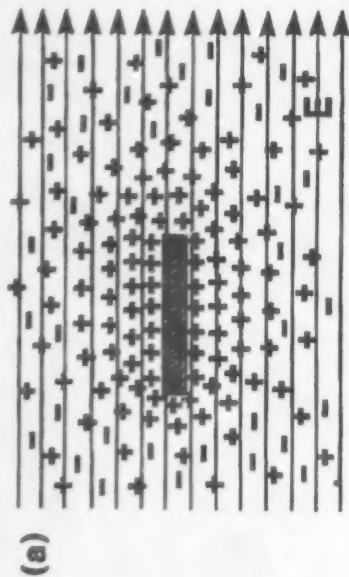
Dependences of the specific surface S_v on average grain diameter d_{gr} (e) and clay content k_{clay} (b) (after O.A. Chernilov and A.I. Kurenkov).



Influence of pH on the cation exchange capacity (a measure of the negative charge) of montmorillonite and humus. Note that below pH 6.0 the charge for the clay mineral is relatively constant. This charge is considered permanent and is due to ionic substitution in the crystal unit. Above pH 6.0 the charge on the mineral colloid increases because of ionization of hydrogen from exposed O—H groups at crystal edges. In contrast to the clay, essentially all of the charges on the organic colloid are considered pH dependent. [Montmorillonite data from Coleman and Mellich (1); organic colloid data from Helling et al. (2).] (Brady, 1974)



Time constant of relaxation for diffusion-limited processes versus radius of particle. The "C" are data from Collet (1959), and the "G" are data from Grisseman (1971) after Pelton et al. (1978). The line labeled "W" is predicted by the Wong (1979) model and that labeled "T" is predicted by the Trukhin (1962) model (see Pethig, 1979). The dashed line is time constant = (radius)²/3.4 × 10⁻⁶ s. (Olhoeft, 1985)

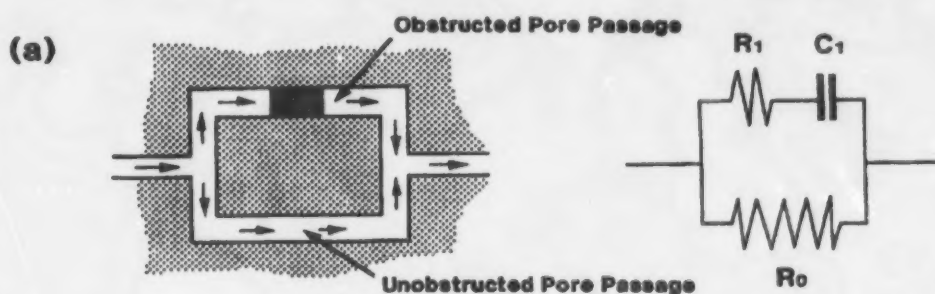


Development of membrane polarization associated with a constriction within a channel between mineral grains, and (B) negatively charged clay particles (Frammer et al. 1994) and fibrous elements along the sides of a channel

(Reynolds, 1997)

CONCEPTUAL SOIL MODEL

EQUIVALENT CIRCUIT



Non-Conducting Soil Matrix.



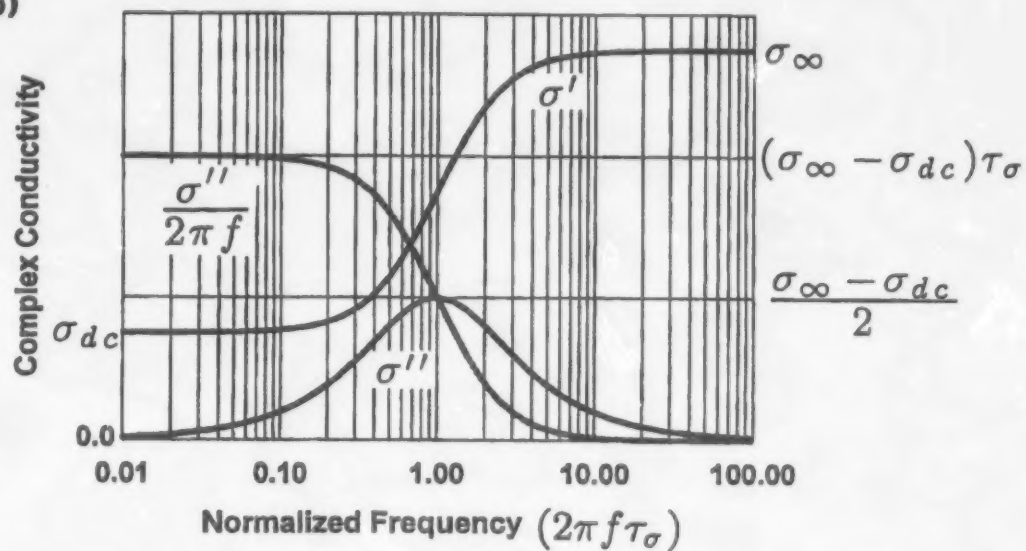
Electrolytic Charge Transport.

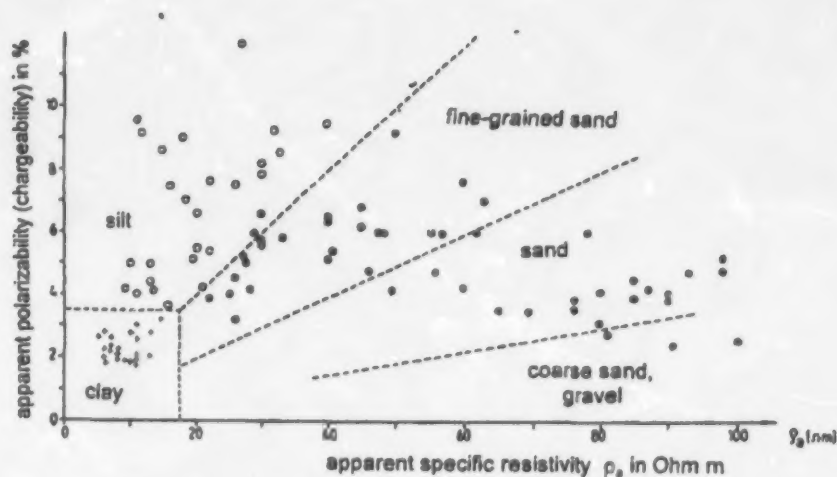
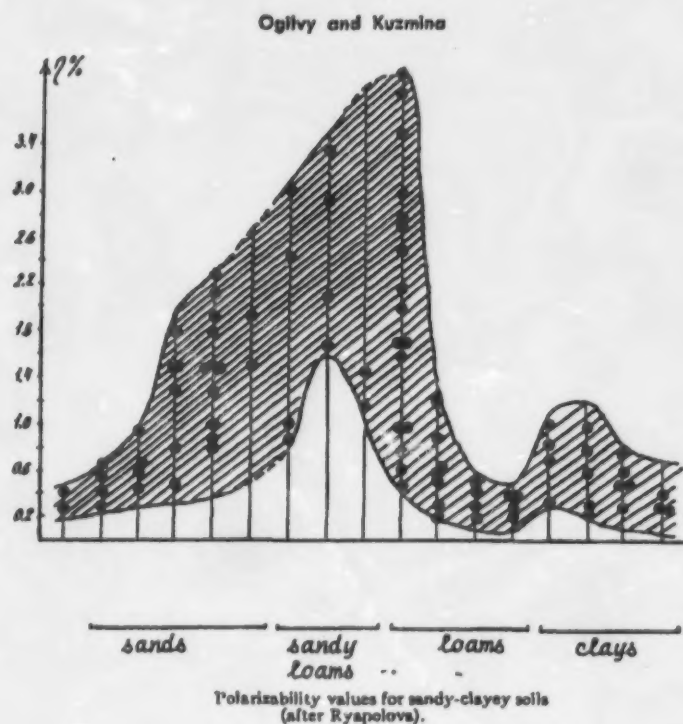
R_0 - Effective Resistance Unobstructed Pore Passage.

R_1 - Effective Resistance Obstructed Pore Passage.

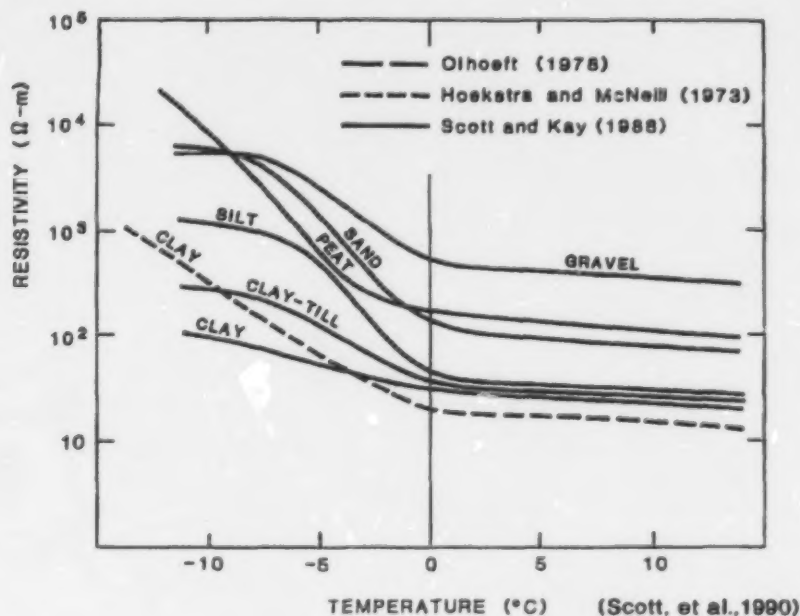
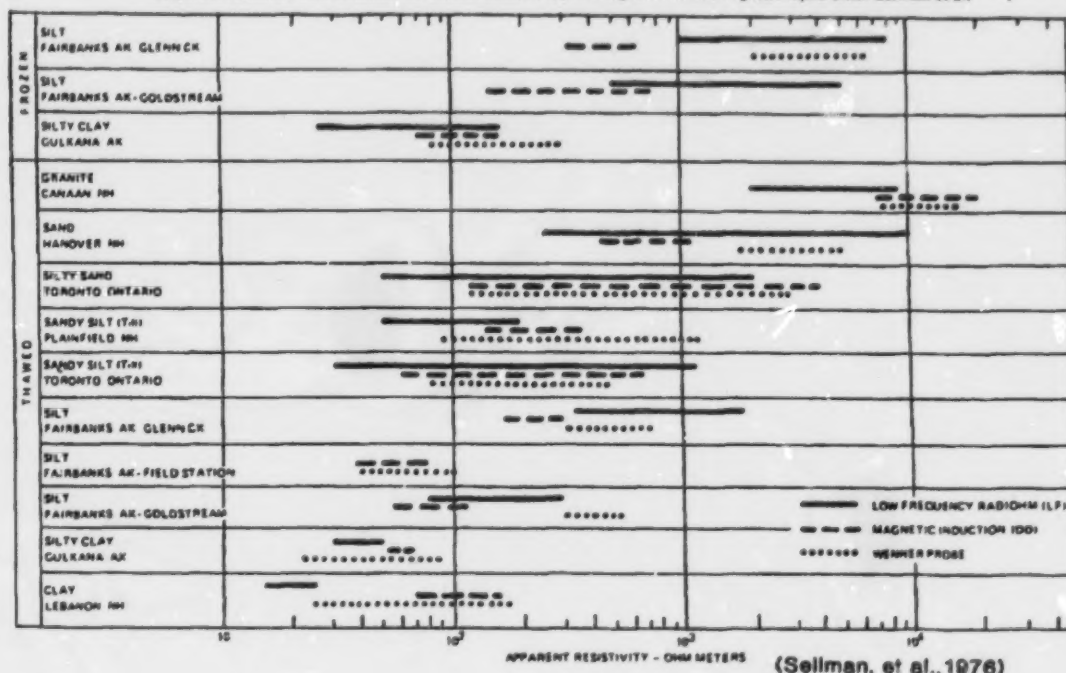
C_1 - Effective Capacitance Obstructed Pore Passage.

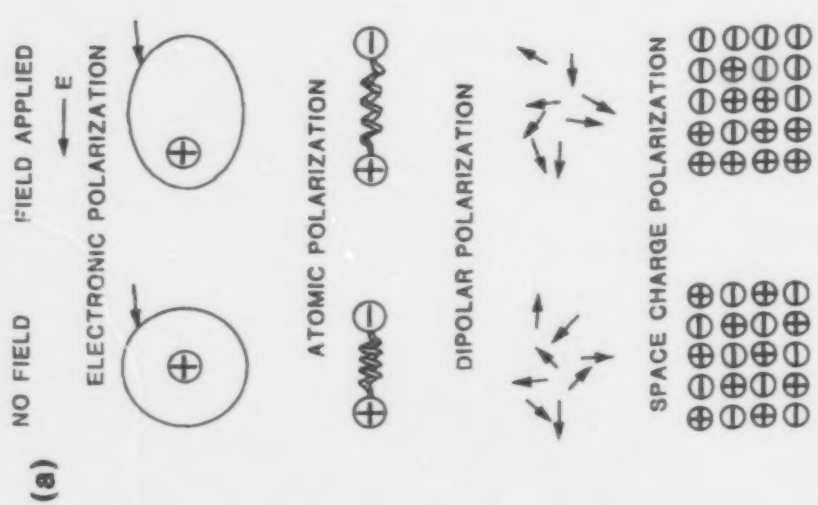
(b)



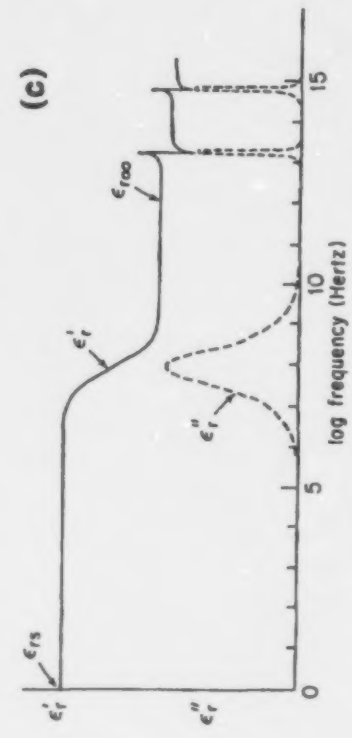
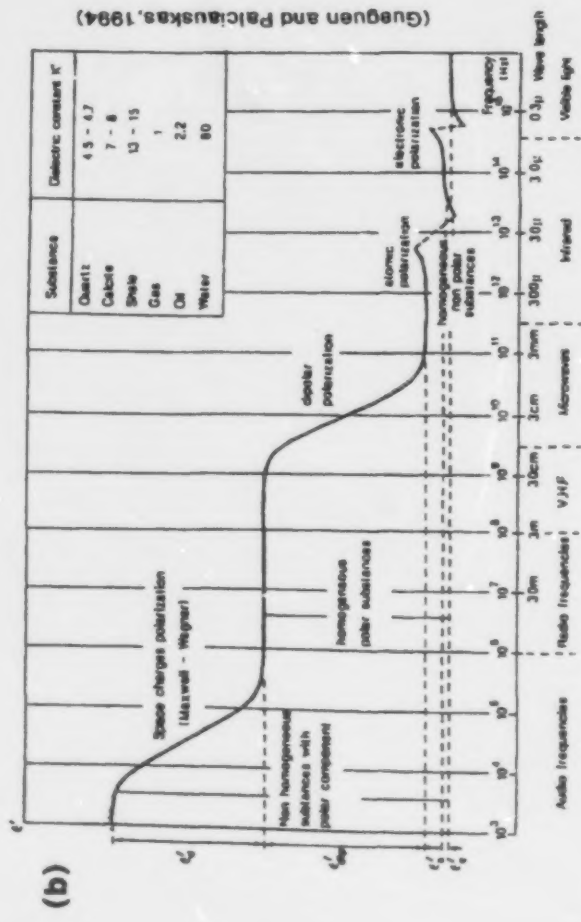


Apparent resistivity data obtained for various material types using three measuring techniques (after: Selman et al., 1976)



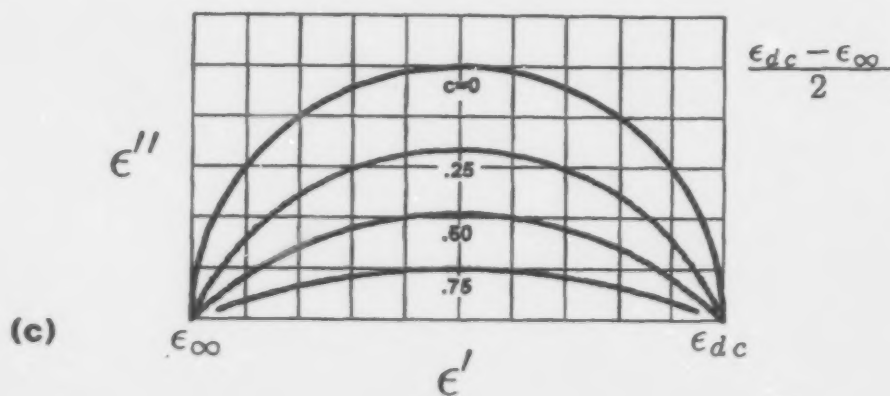
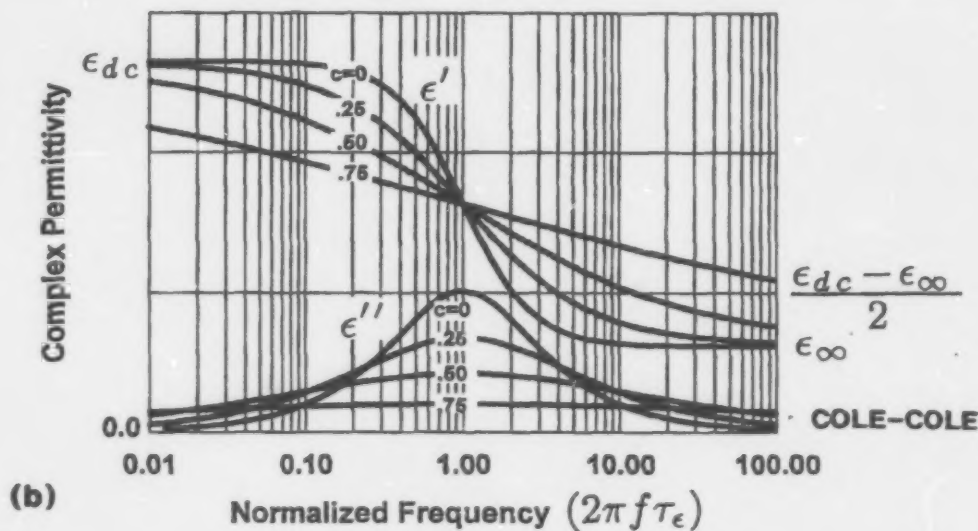
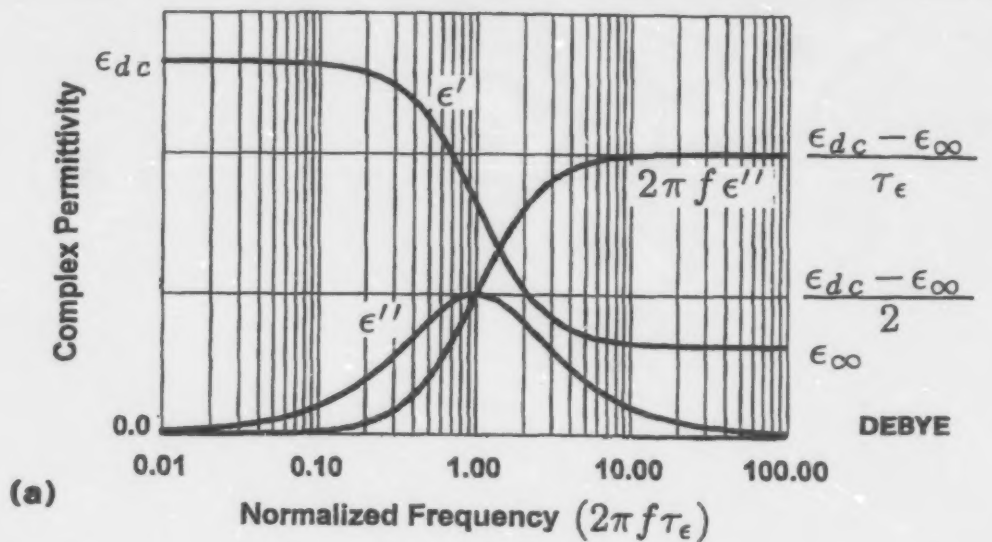


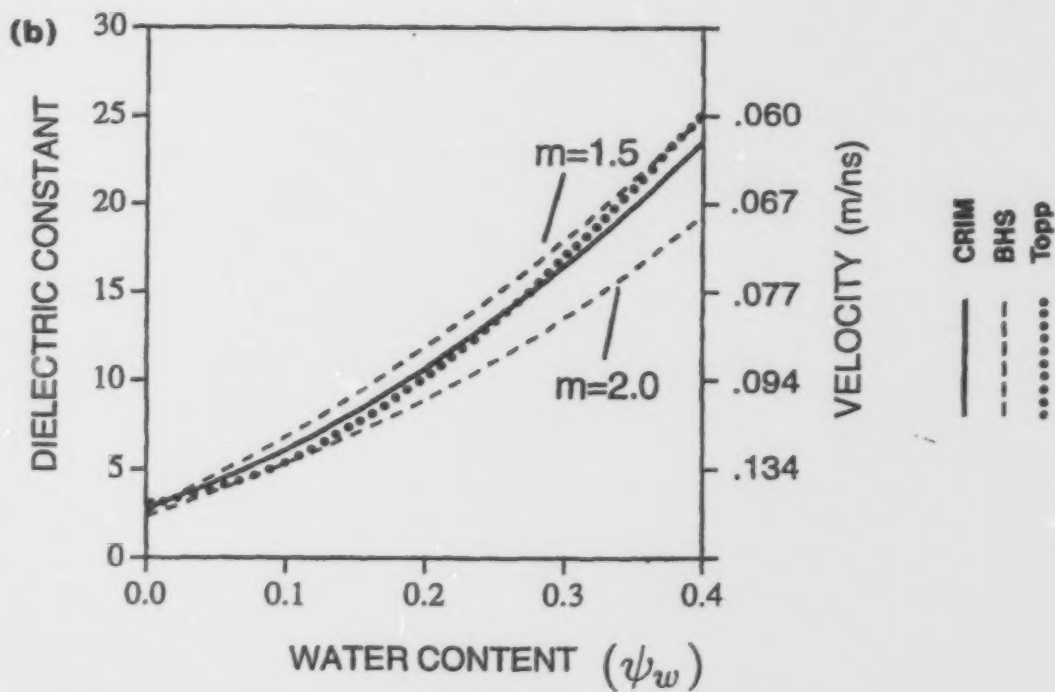
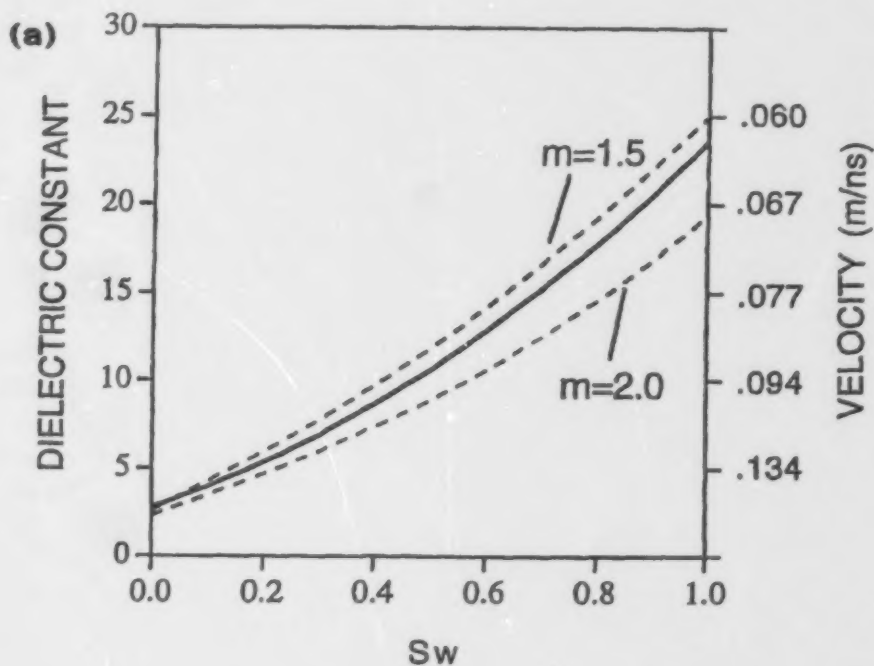
Mechanisms for polarization (from von Hippel 1954 a, b).
(von Hippel, 1954)



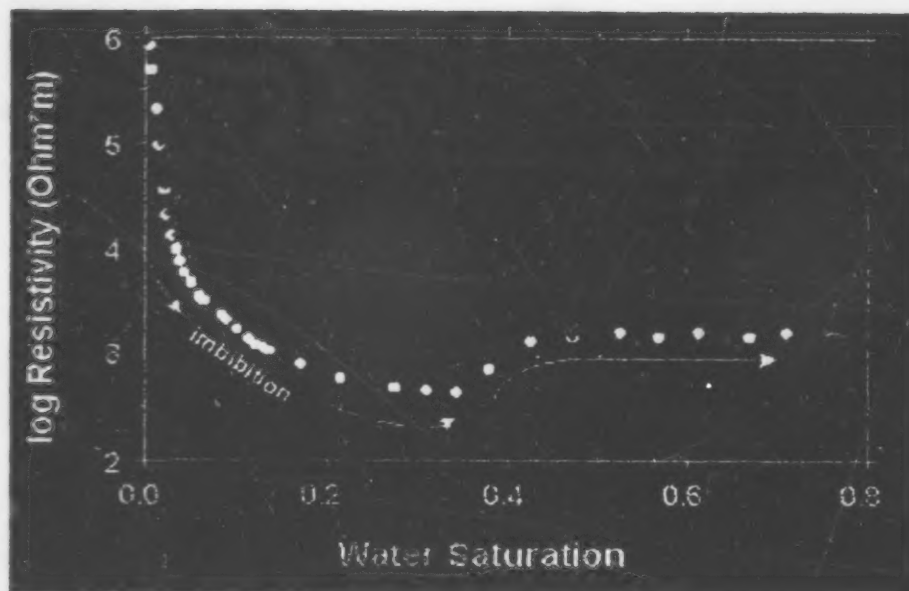
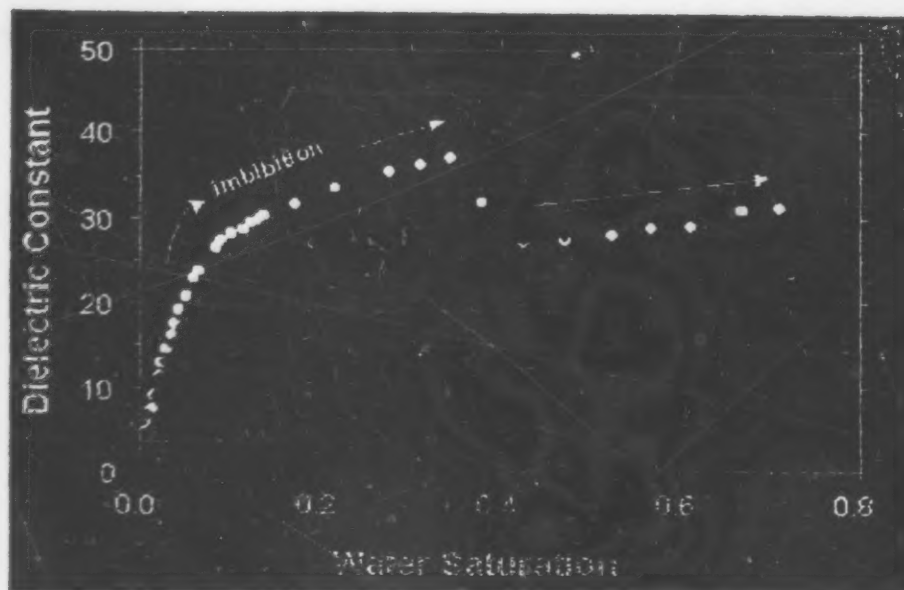
Schematic diagram of relaxation region, atomic and electronic resonance regions in $\epsilon_r = \epsilon' - j\epsilon''$.
(King and Smith, 1981)

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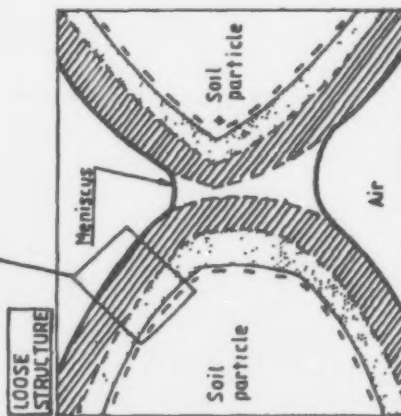
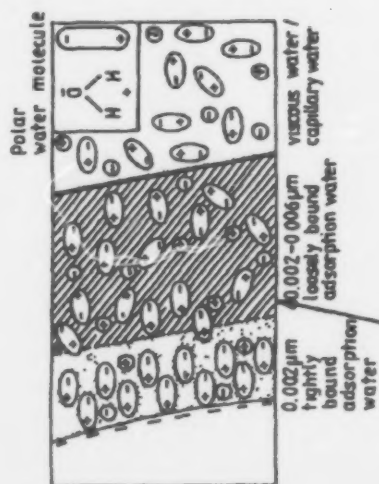
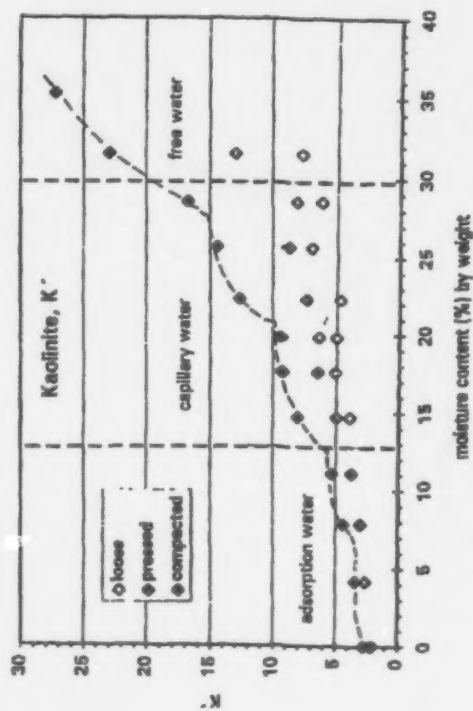
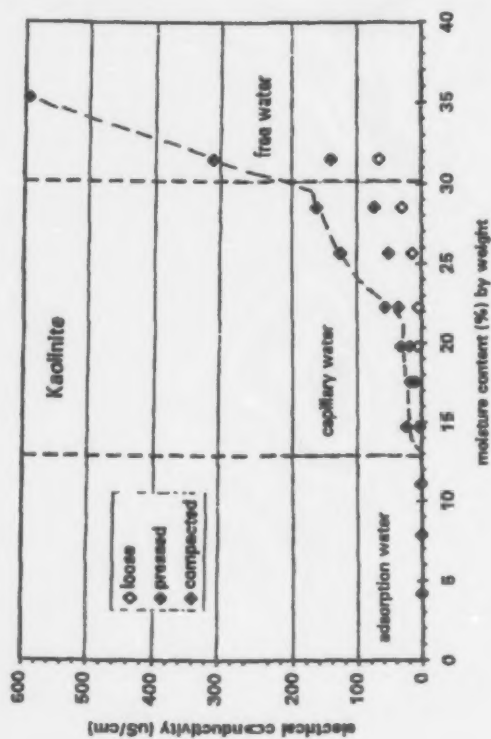




Modified from Greaves et al.(1996)



(Knight et al., 1999)



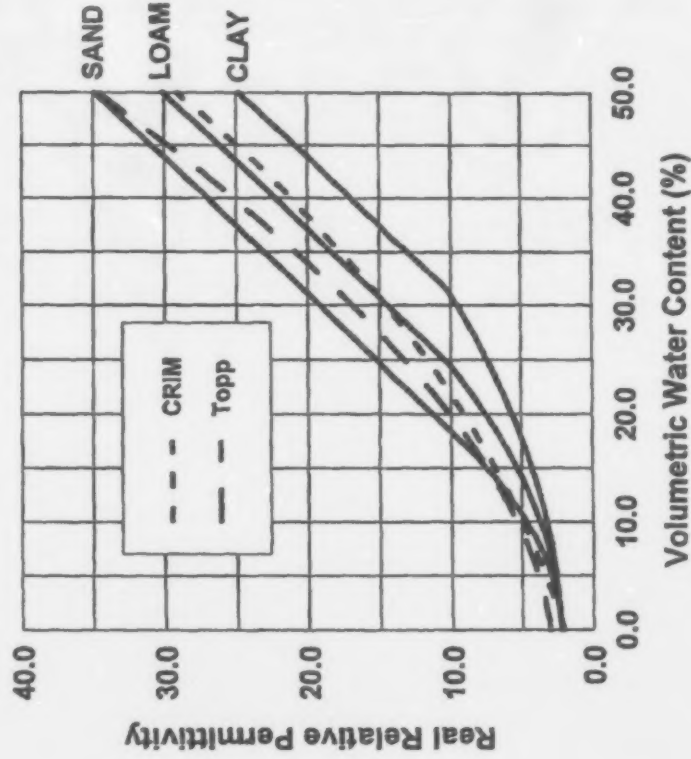
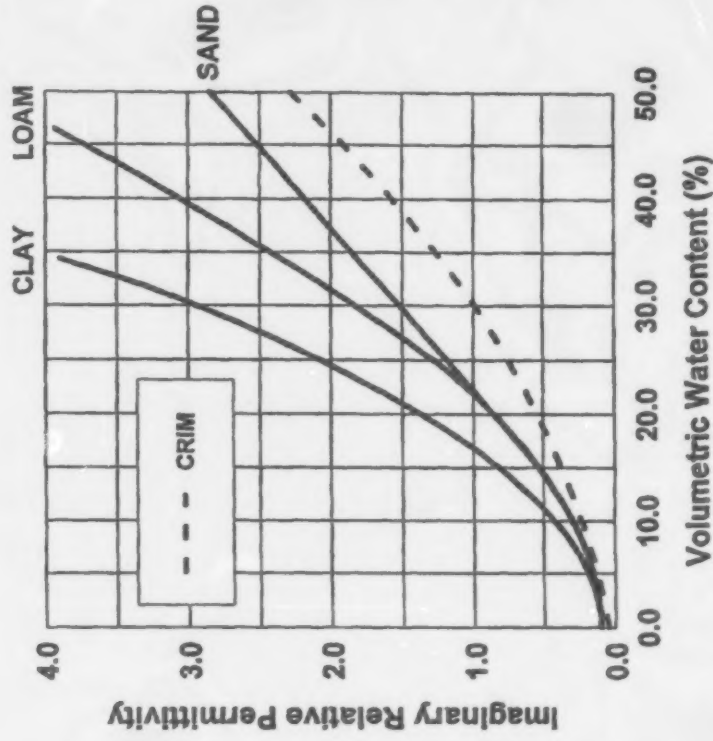
(Saarenketo, 1998)

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SOIL ELECTRICAL PROPERTIES - MOISTURE

Figure

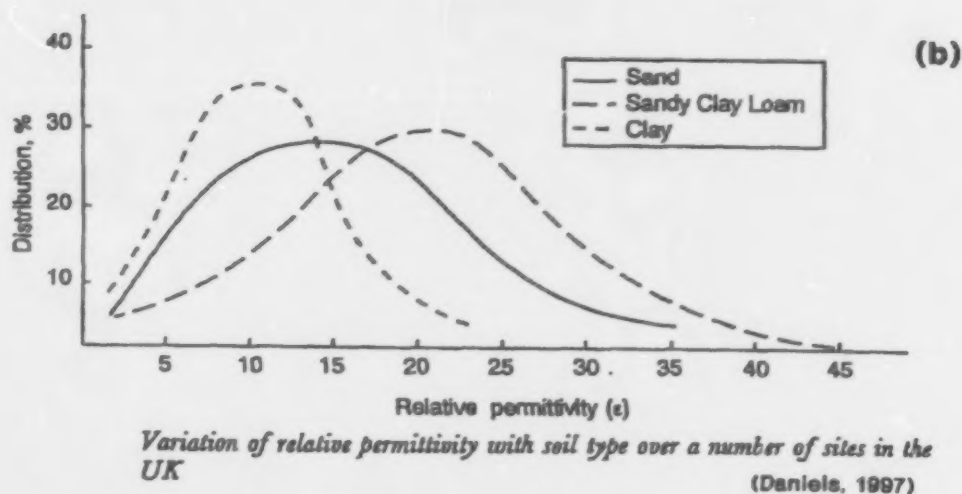
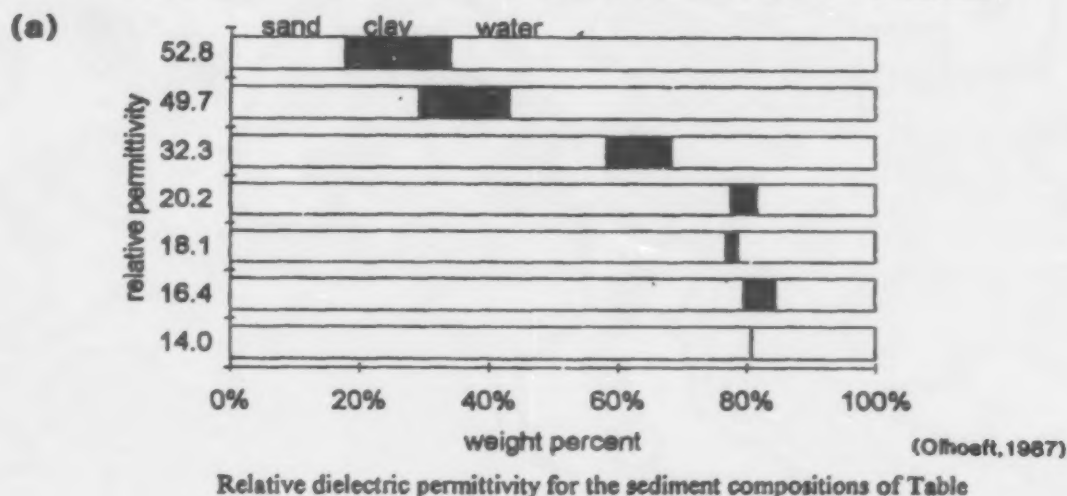


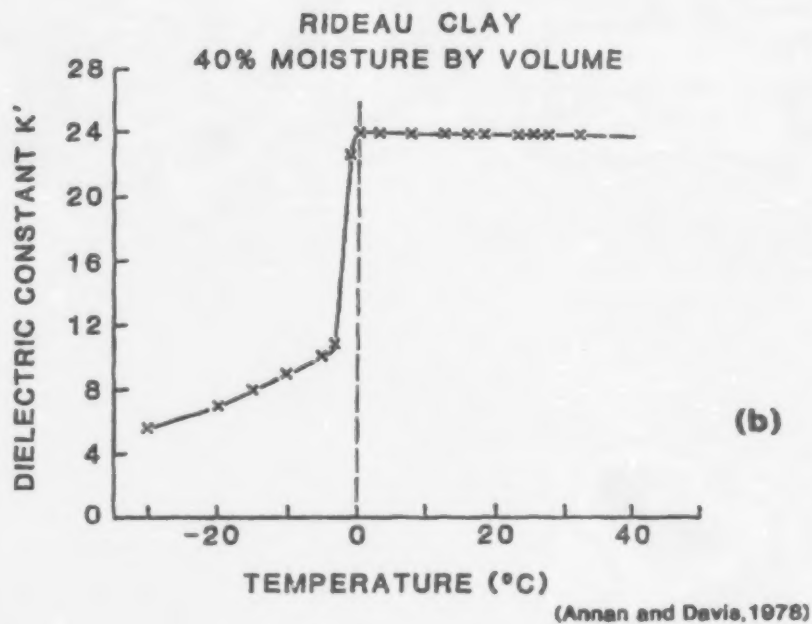
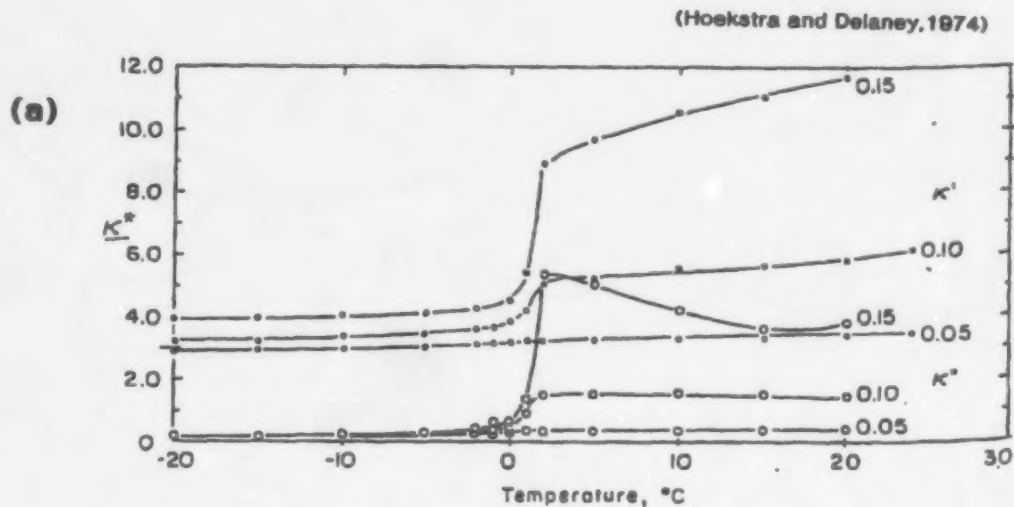
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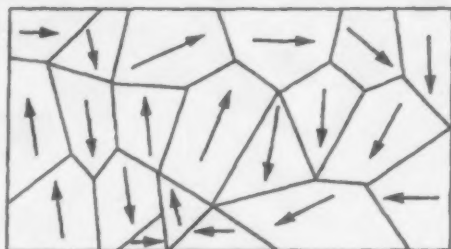
PERMITTIVITY MOISTURE MODELS

Sand	Clay weight - %	Water /	Relative permittivity	Loss tangent	Specific resistivity
83.6	0	16.4	12.1	0.088	269
80.7	0	19.3	14.0	0.14	125
79.5	4.9	15.6	16.4	0.91	15.7
76.7	1.9	21.4	18.1	0.40	33.5
77.5	4.0	18.5	20.2	0.84	13.4
58.2	10.1	31.7	32.3	1.09	6.6
29.3	13.9	56.8	49.7	1.00	4.7
17.9	16.5	66.4	52.8	1.00	4.4
0	22.9	77.1	62.6	1.14	3.4





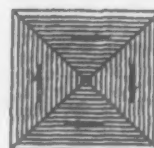
(Annan and Davis, 1978)



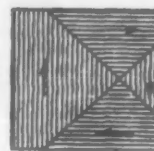
(a)

Domains in a ferromagnetic crystal. Inside each domain, moments are aligned parallel to each other. Domain size is typically $10^{-2} - 1 \mu\text{m}$.

(Sears and Zemansky, 1964)

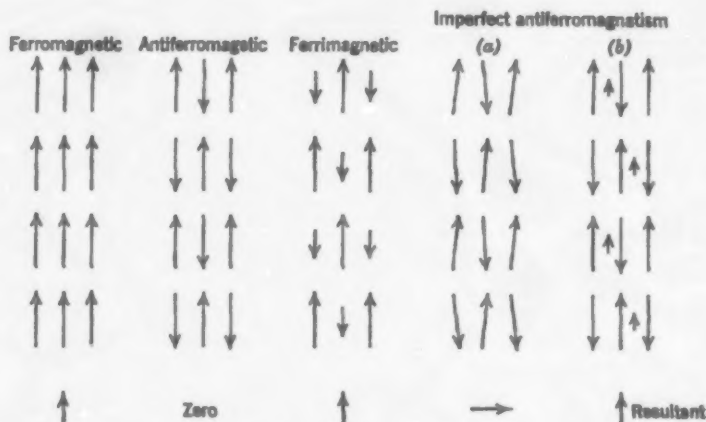


Unmagnetized

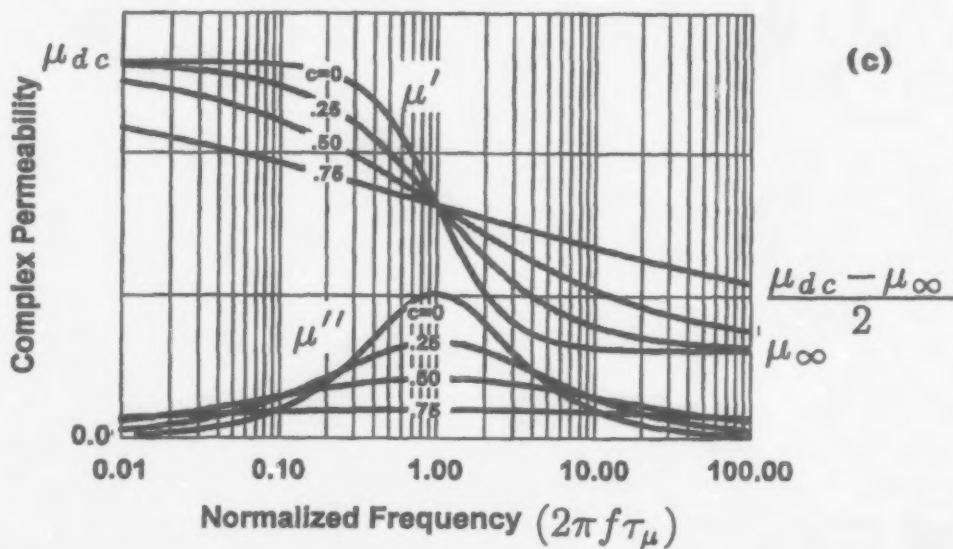


Magnetized by domain growth (boundary displacement)

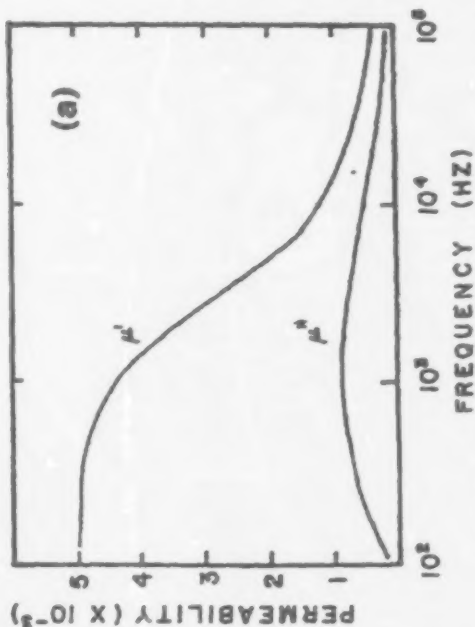
(b)



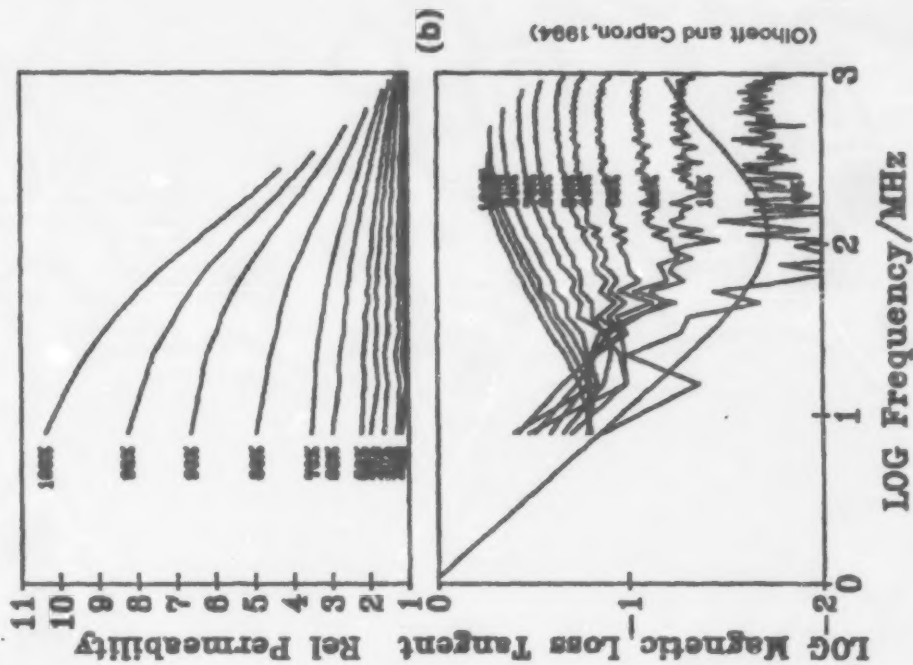
Schematic representation of spontaneous magnetisation in crystals. The arrows represent the elementary moments, the resultants being given at the bottom. (Irving, 1964)



(c)



Galt (1952) single crystal magnetite permeability spectrum.



(Olhoeft and Capron, 1984)

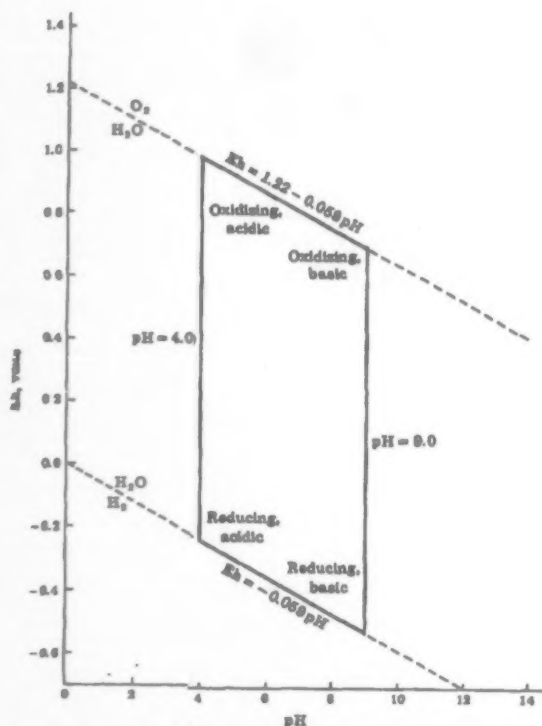
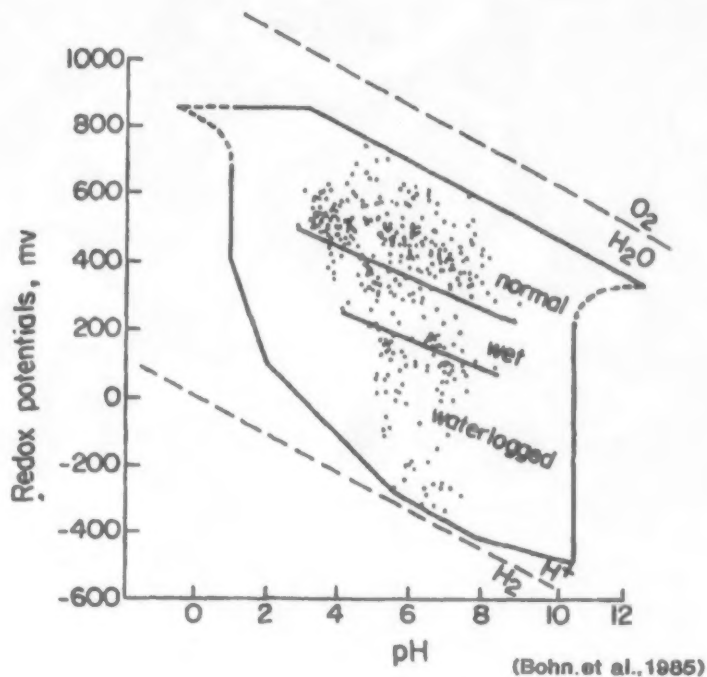
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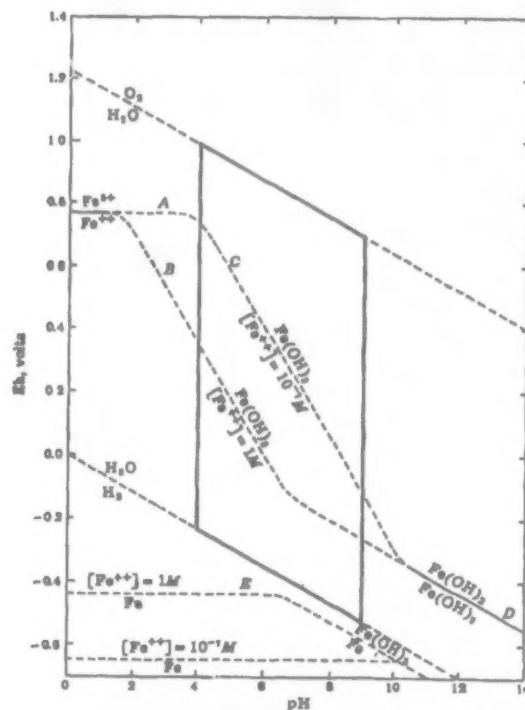
MAGNETIC PERMEABILITY DISPERSION

Figure

The range of redox potentials and pH values in soils. (From L. G. M. Baas Becking, I. R. Kaplan, and D. Moore, *J. Geology* 68:243, (1960).)



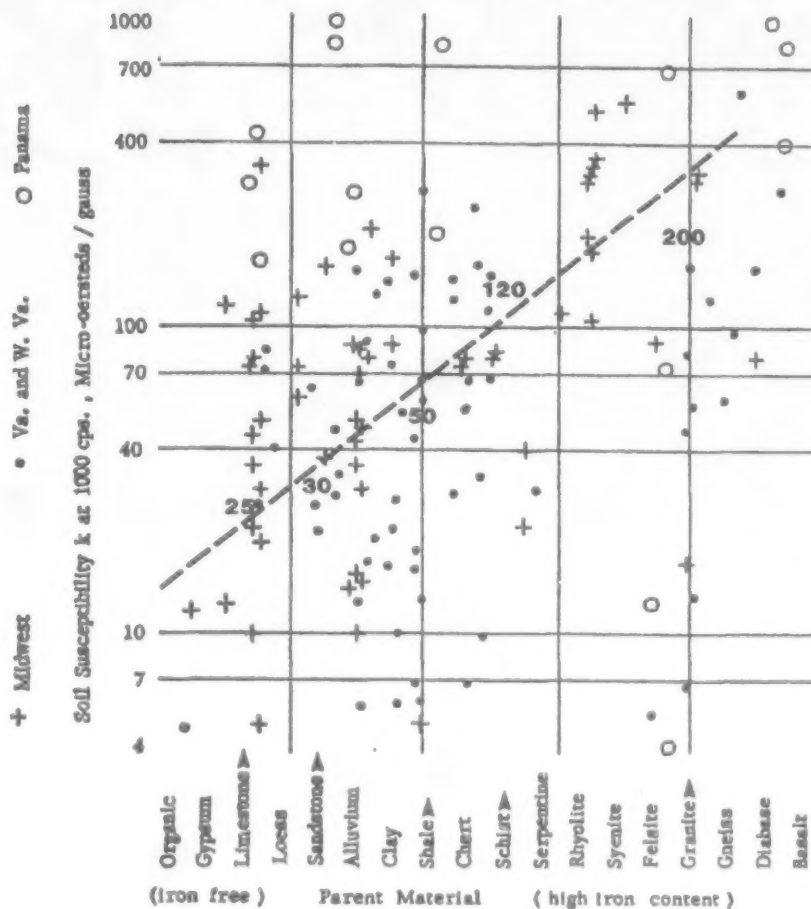
Framework of Eh-pH diagrams. The parallelogram outlines the usual limits of Eh and pH found in near-surface environments.



Eh-pH diagram for the simple ions and hydroxides of iron at 25°C.

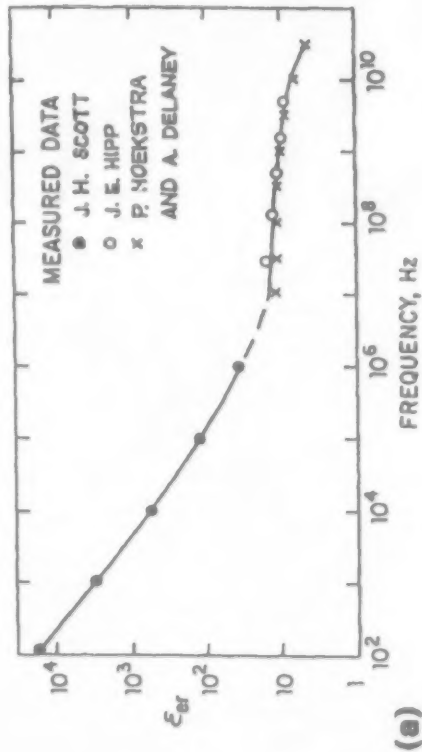
(Krauskopf, 1967)

MAGNETIC PROPERTIES OF TOPSOILS

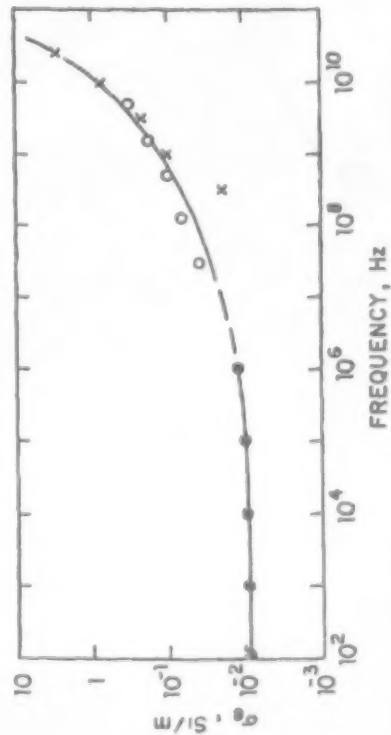


Modified: Cook and Carls (1962)

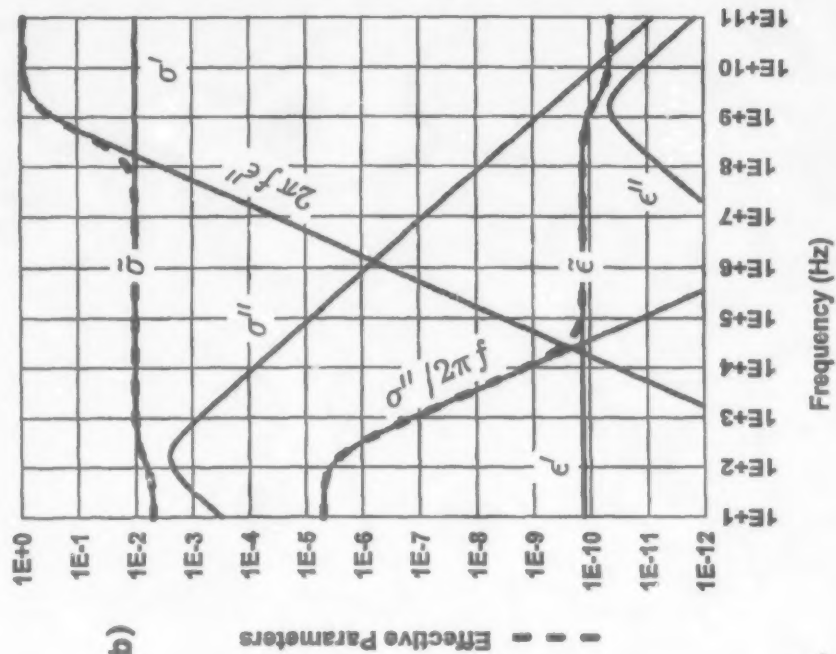
25 Average Parent Rock Susceptibility
(Telford et al., 1976)



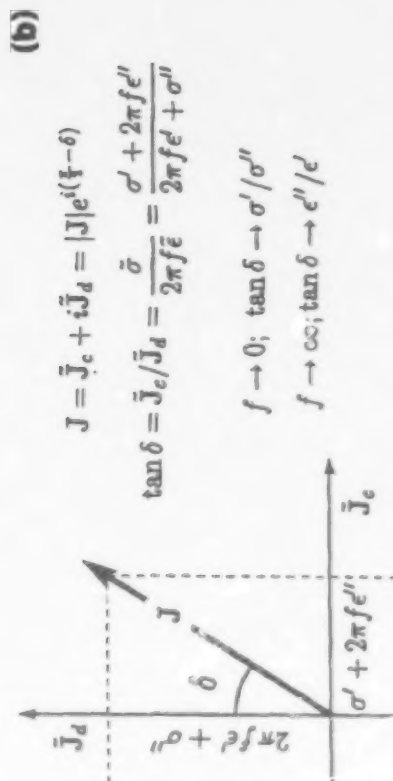
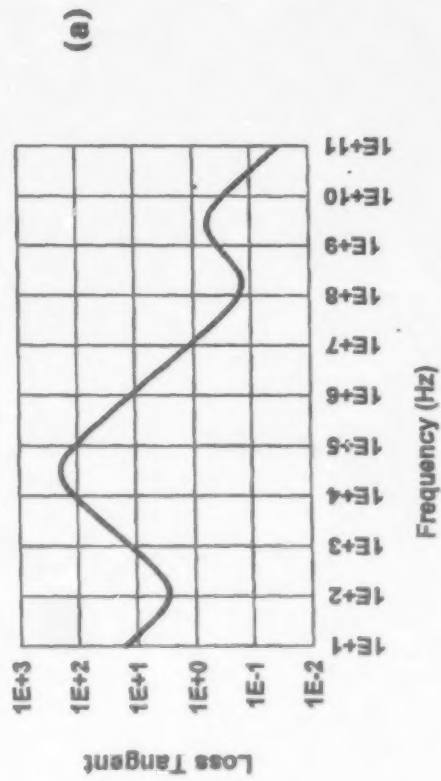
(a)

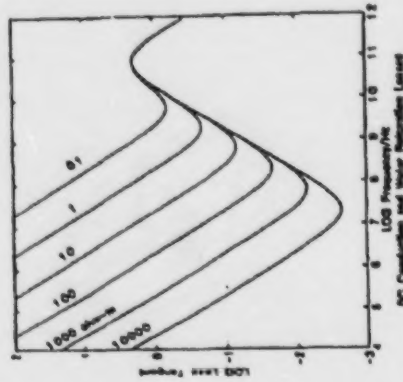


(b)

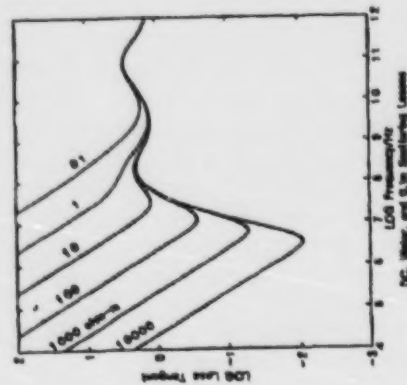


Relative effective permittivity ϵ_{er} and effective conductivity σ_e as a function of frequency for a typical clay-loam soil with a water content of about 10 % by weight.
 (King and Smith, 1981)

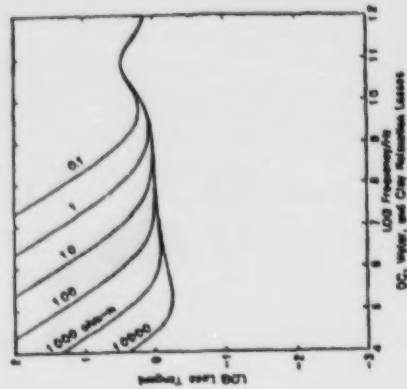




Log loss tangent versus log frequency for EM losses of water at several electrical resistivities. The "peak" in the loss tangent that all the curves converge on at high frequencies is the dielectric relaxation loss due to the rotational polarization of the water molecule. For homogeneous, clay-free, wet soils or rocks, the loss tangent would be the same but with the water relaxation peak reduced in proportion to the reduced amount of water.

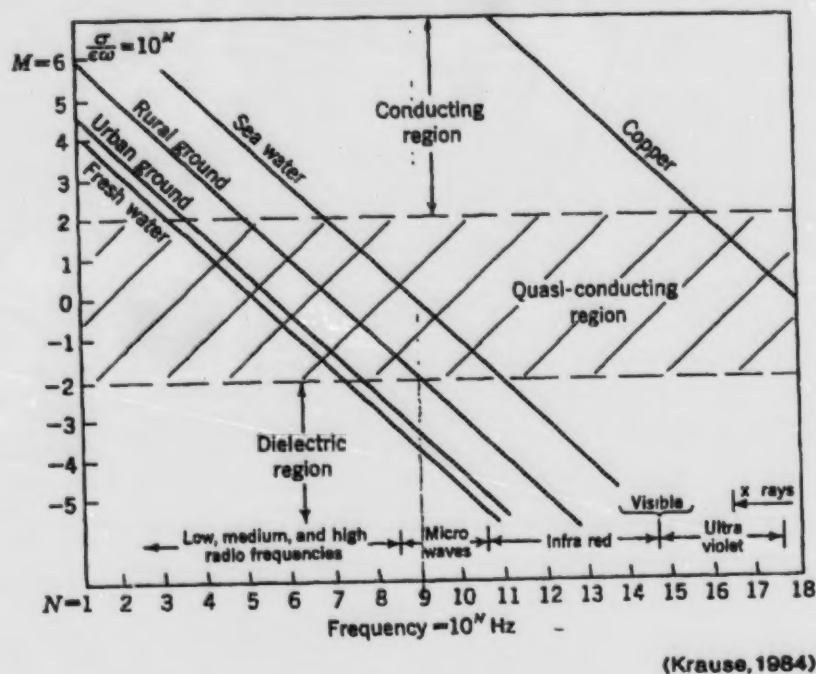
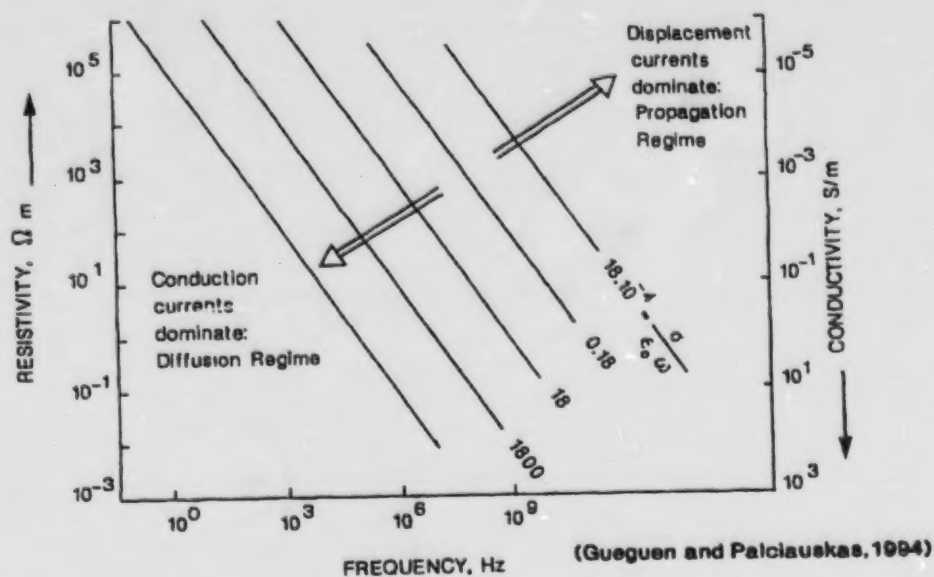


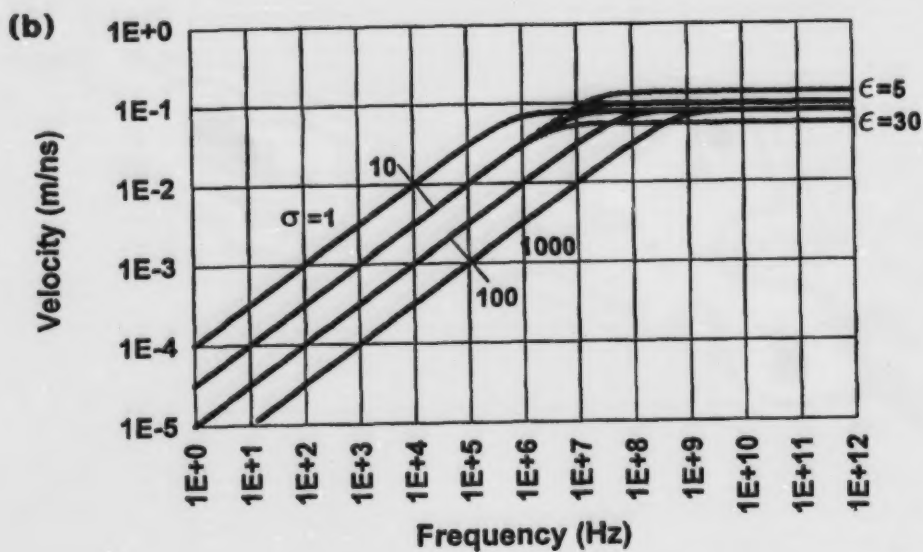
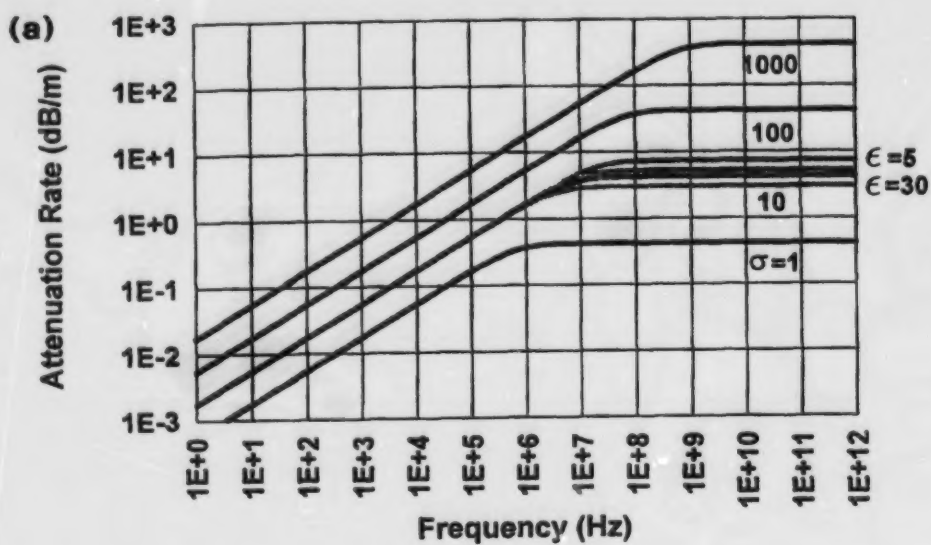
EM losses from Figure 1 with the addition of scattering from a log-normal distribution of spherical particles with a mean diameter of 0.1 m and a standard deviation of 1 m.



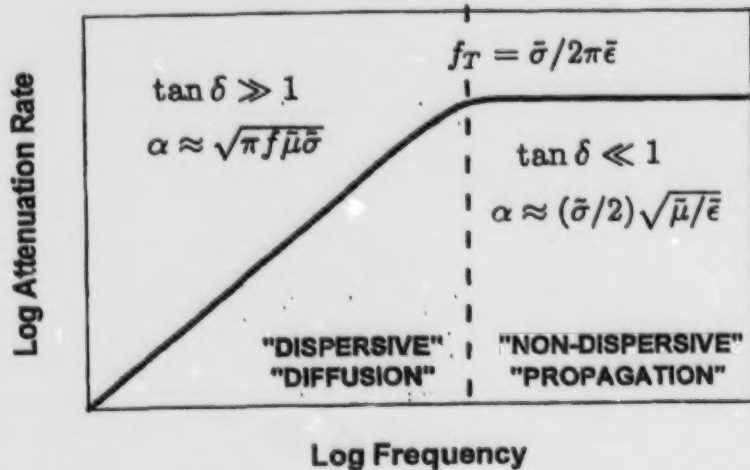
EM losses from Figure 1 with the addition of diffusion-limited dielectric relaxation from colloidal size particles. The 1 Ω -m curve is typical of montmorillonite clay.

(Olhoeft, 1984)

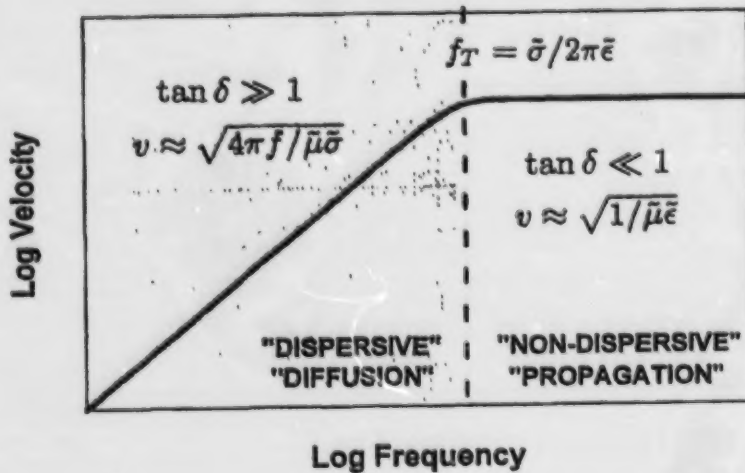




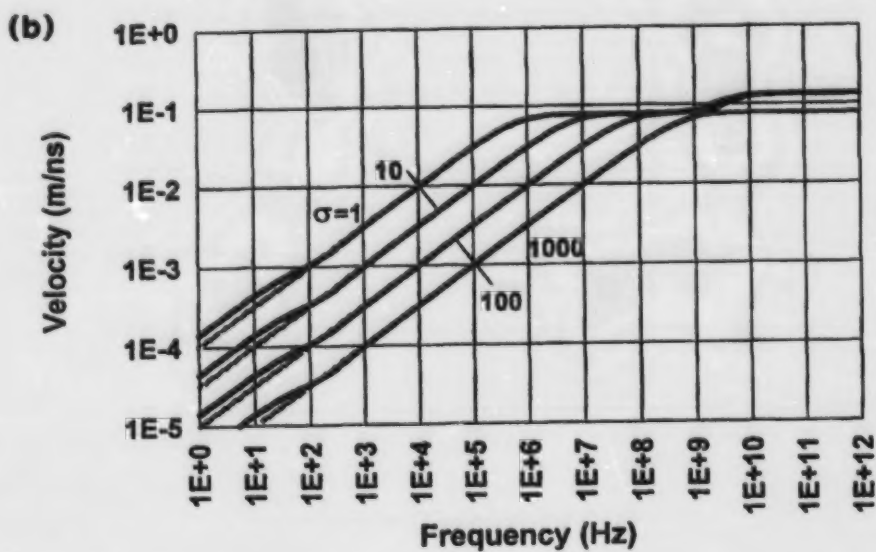
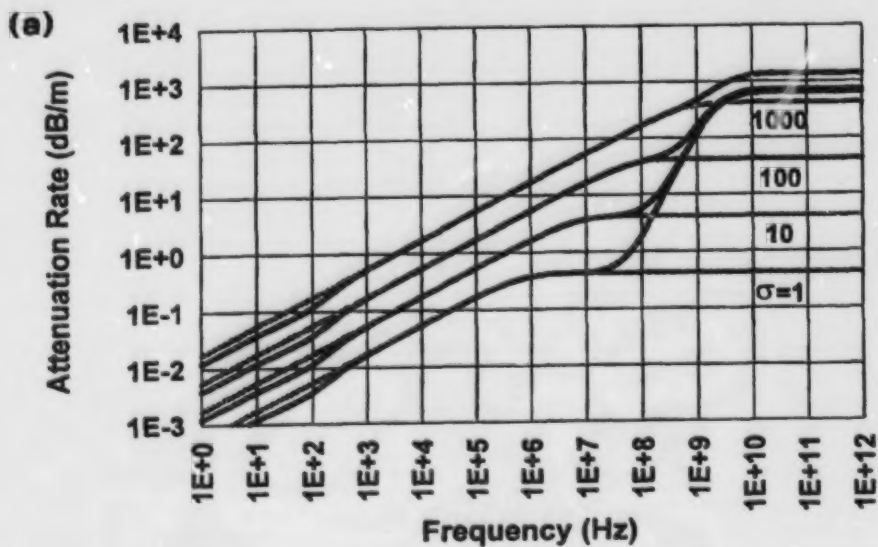
(a)

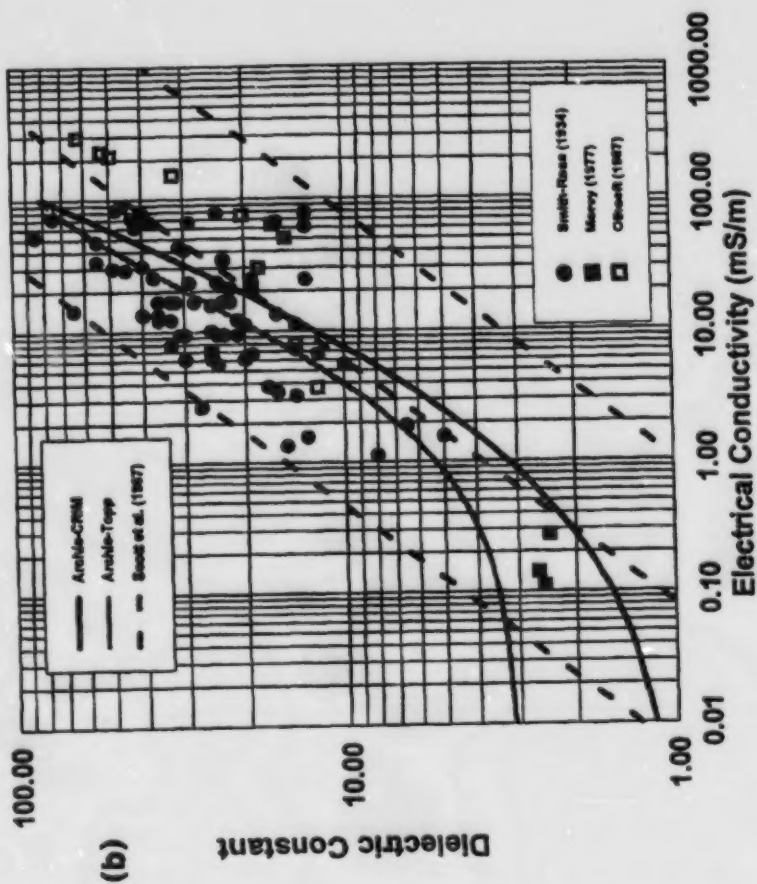
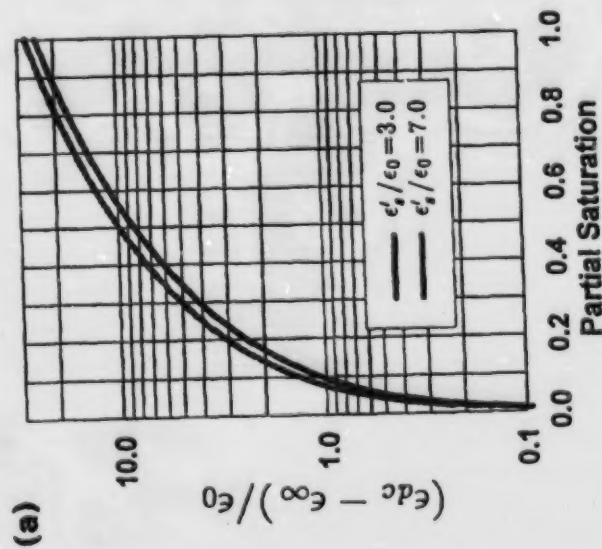


(b)



After: Annan (1992)

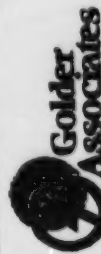




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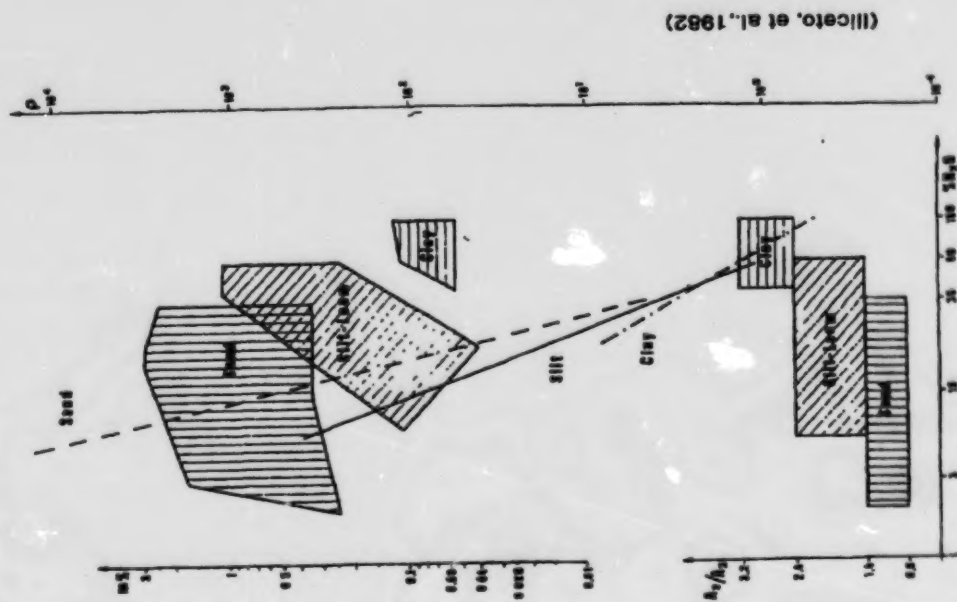


ELECTRICAL CONDUCTIVITY - PERMITTIVITY

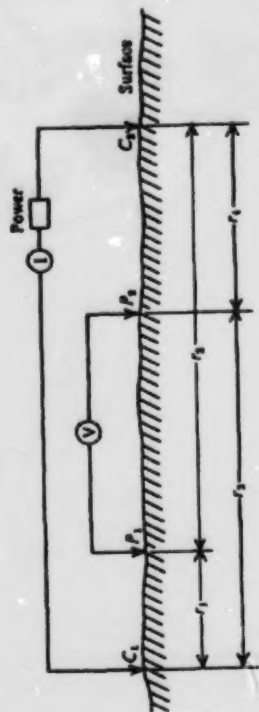
Figure

34

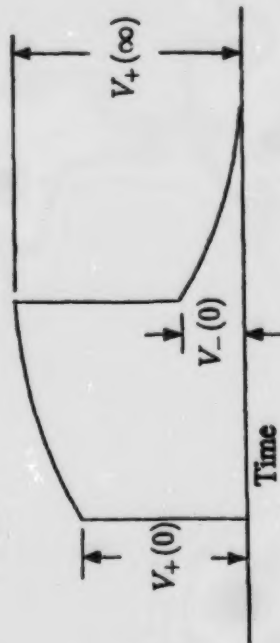
IDENTIFICATION OF SEDIMENTS



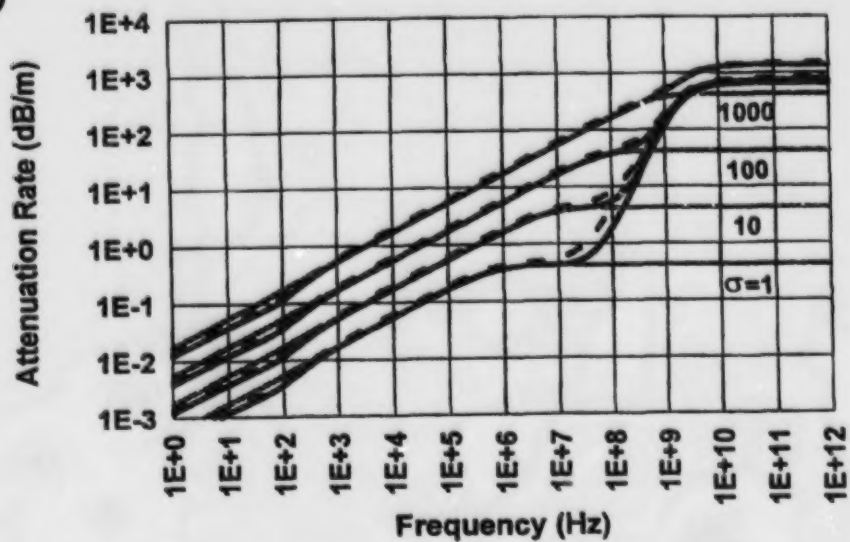
Chargeability a_1 , ratio A_1/A_2 , and electrical resistivity [ρ_w] versus water content in sands, silts, loams and clays.



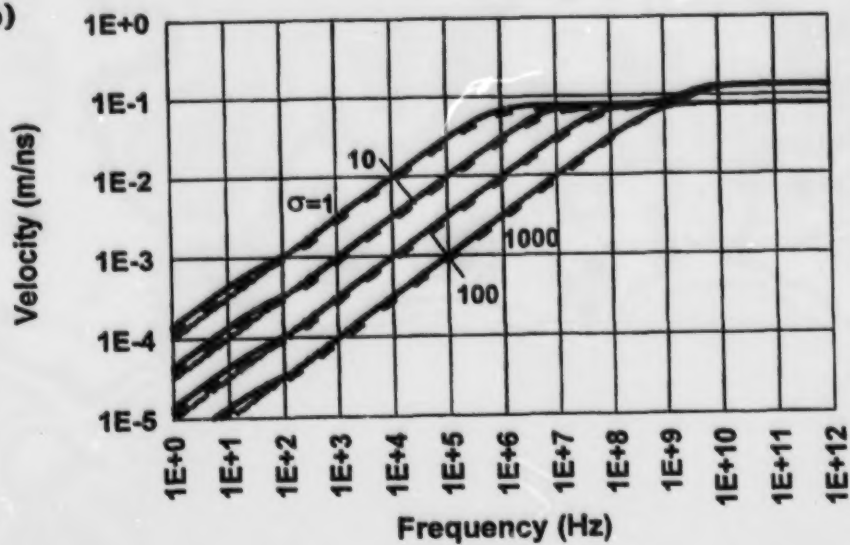
Two current and two potential electrodes on the surface of homogeneous isotropic ground of resistivity ρ .

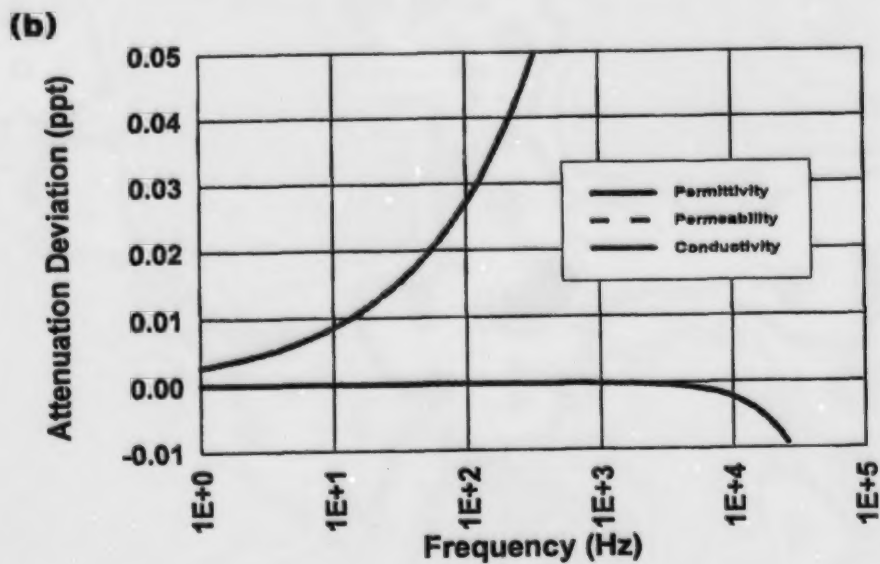
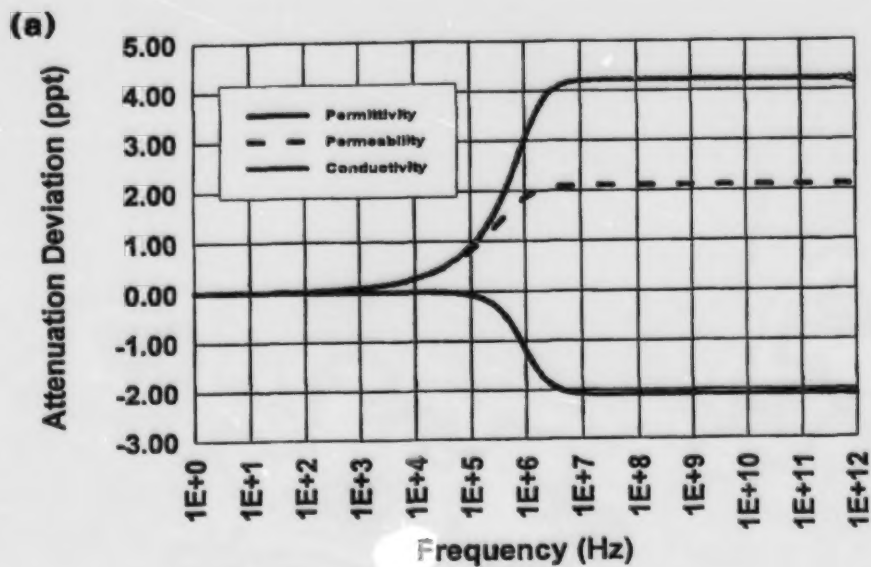


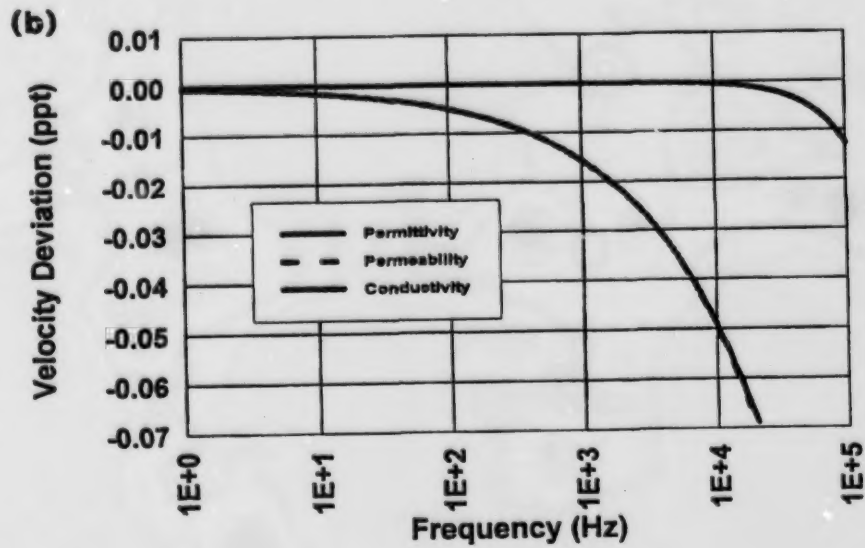
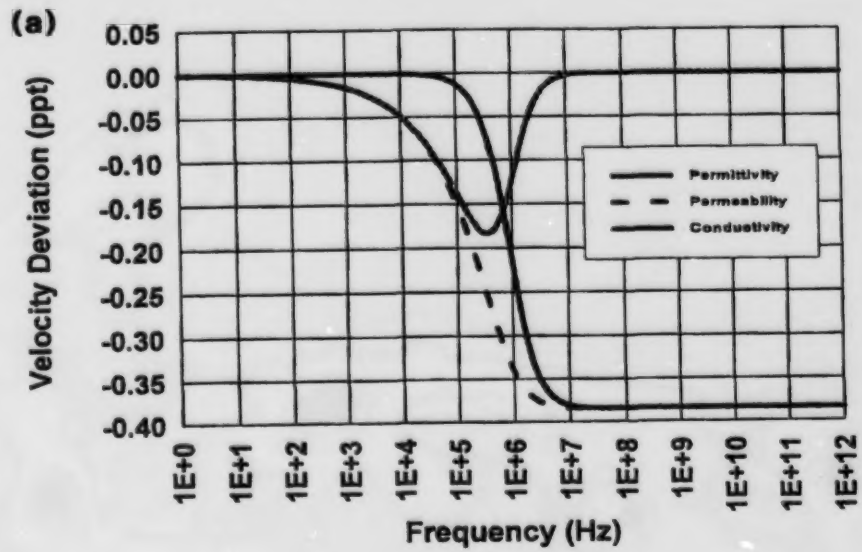
(a)



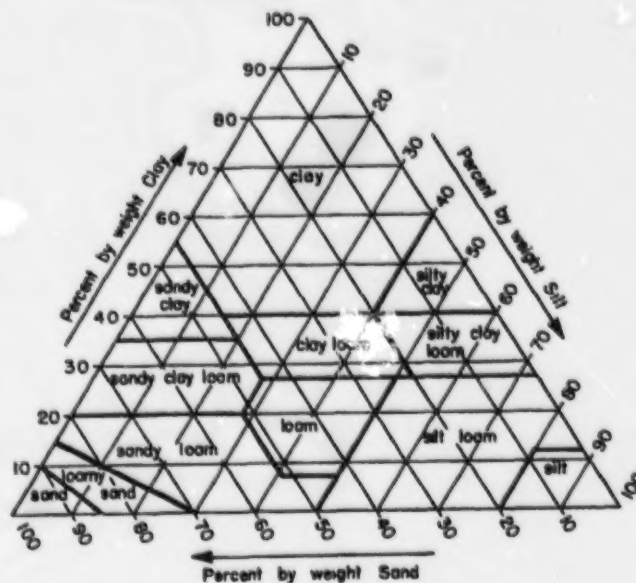
(b)







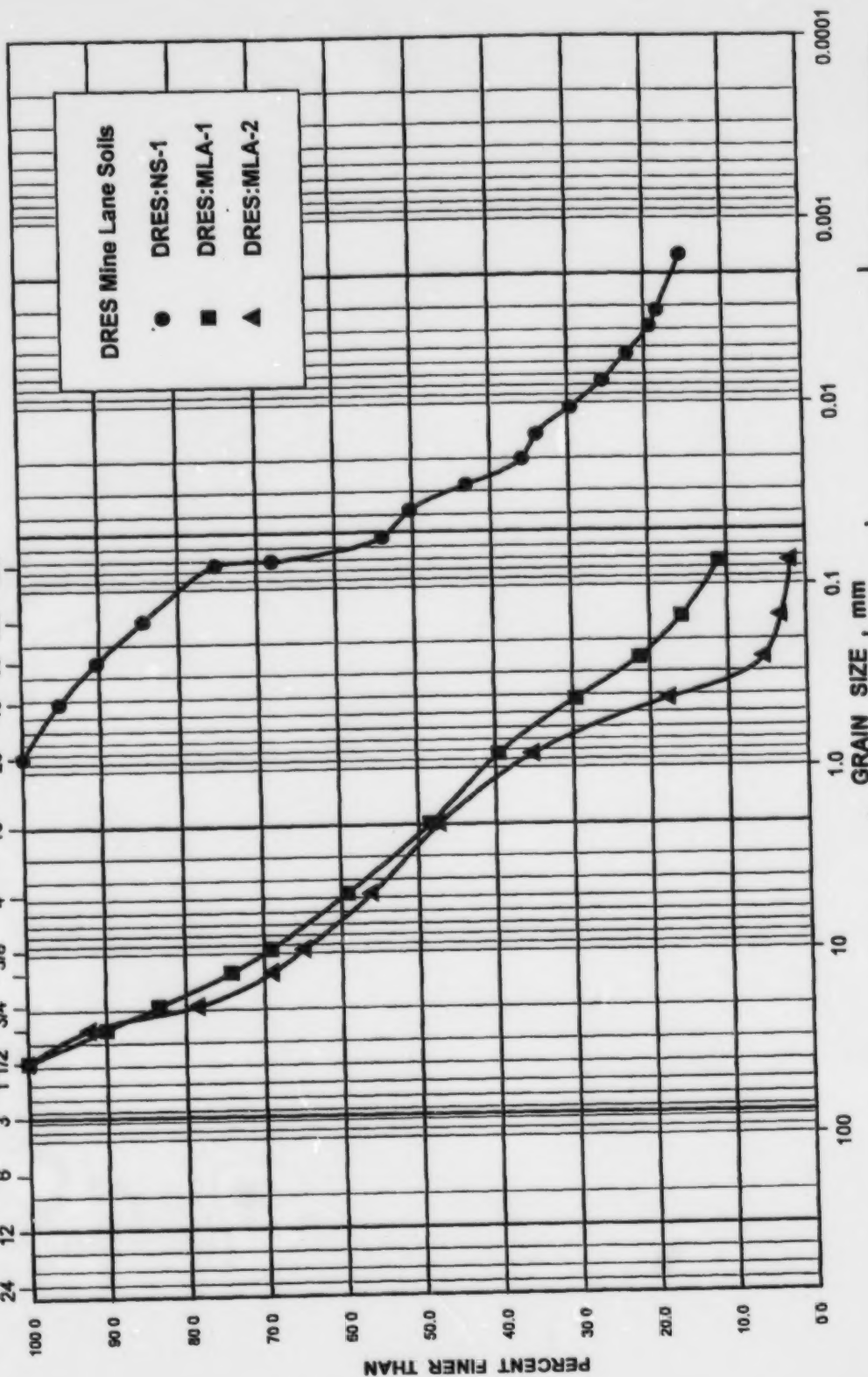
- Textural classification is based on the soil fraction (sample) passing a 3 in. (75.0 mm) sieve. The percentage of cobbles (75.0-200.0 mm) and the percentage of boulders (> 200.0 mm) comprising the larger soil volume are estimated visually.
- On the basis of grain size analysis by standard sieve and/or hydrometer methods, a soil sample comprising a wide range of differing particle sizes with no dominant size fraction (continuous gradation) is described as well graded. A soil sample including a predominance of particles of one or more restricted size ranges (discontinuous gradation) is described as poorly graded.
- The gravel fraction (> 2.0 mm) is determined as the percent weight of sample retained on and above the 10 mesh sieve.
- Residual sample composition (percent by weight of sample passing the 10 mesh sieve) is tabulated according to USDA grain size designations: sand (0.05-2.0 mm); silt (0.002-0.05 mm); clay (< 0.002 mm), and classified as per the following USDA textural triangle.



- The soil is assigned the residual sample USDA classification, with gravel content described as tabulated below and with appropriate notation of cobble and boulder content.

Percent Gravel by	Modifier	Example
> 50	gravel and	gravel and loam
35-50	and gravel	loam and gravel
20 - 35	gravely	gravely loam
12 - 20	some gravel	loam - some gravel
5 - 12	little gravel	loam - little gravel
0 - 5	trace gravel	loam - trace gravel

24 12 6 3 1 1/2 3/4 3/8 4 10 20 40 60 100 200



CLAY SIZE

SILT SIZE

SAND SIZE

GRAVEL SIZE

BOULDER SIZE

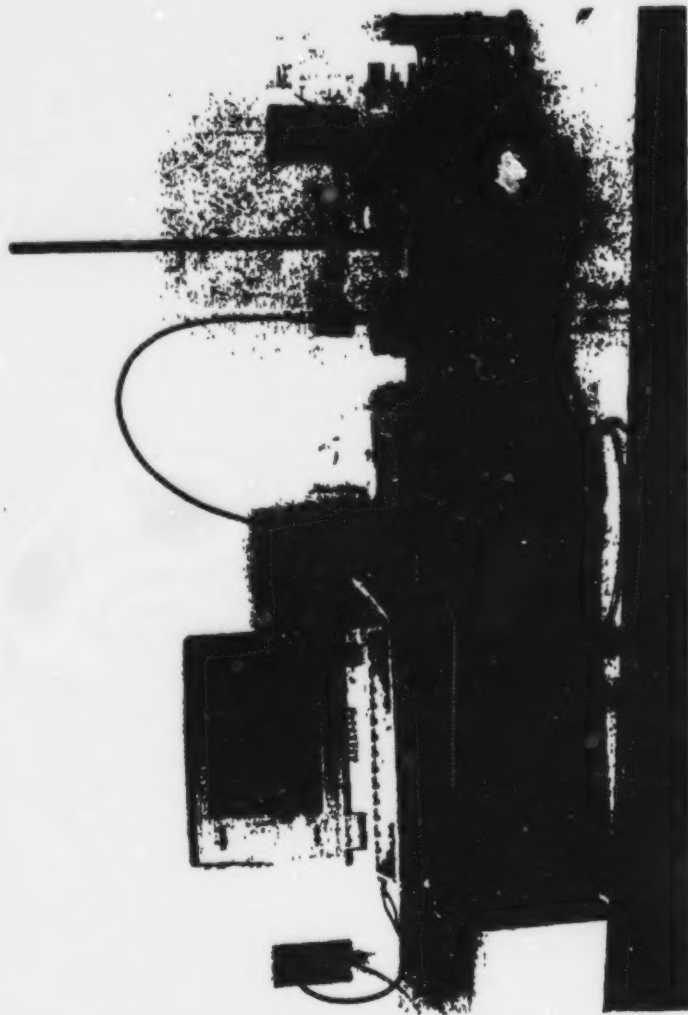
GRAIN SIZE DISTRIBUTION - DRES SOILS

Figure

40

Project No. 9721145
 Drawn JG
 Reviewed WMG
 Date 10/09/98



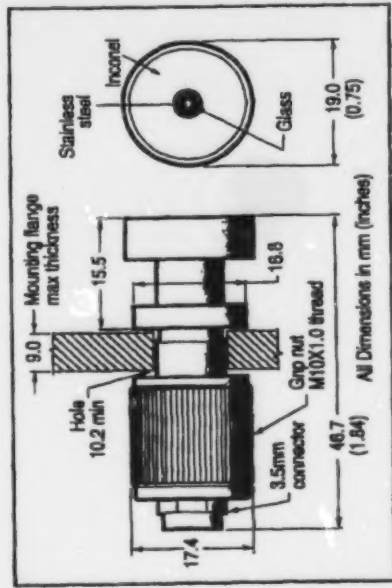
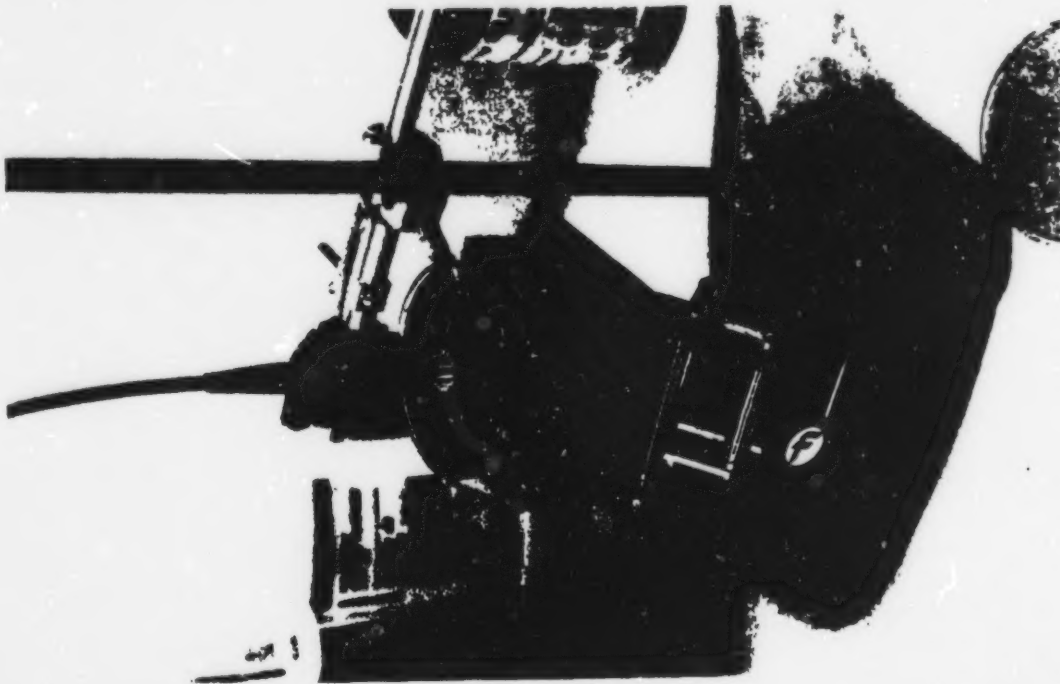


Project No. 972-1145
Drawn GMC
Reviewed
Date 10/98



DIELECTRIC PROBE APPARATUS

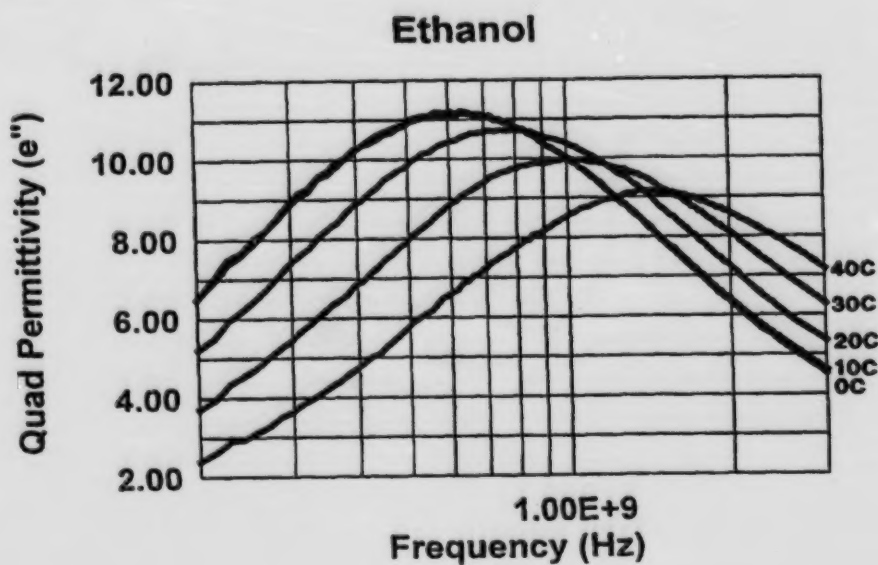
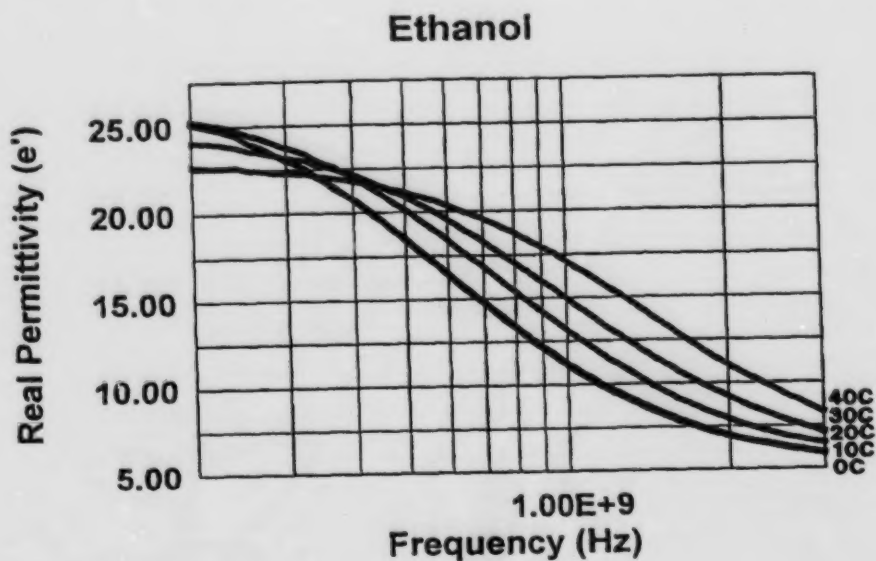
Figure 41



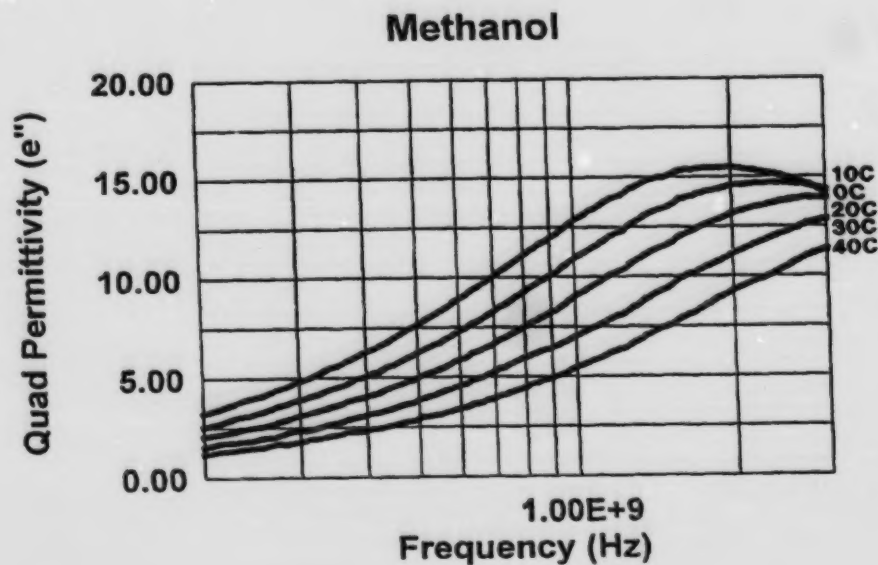
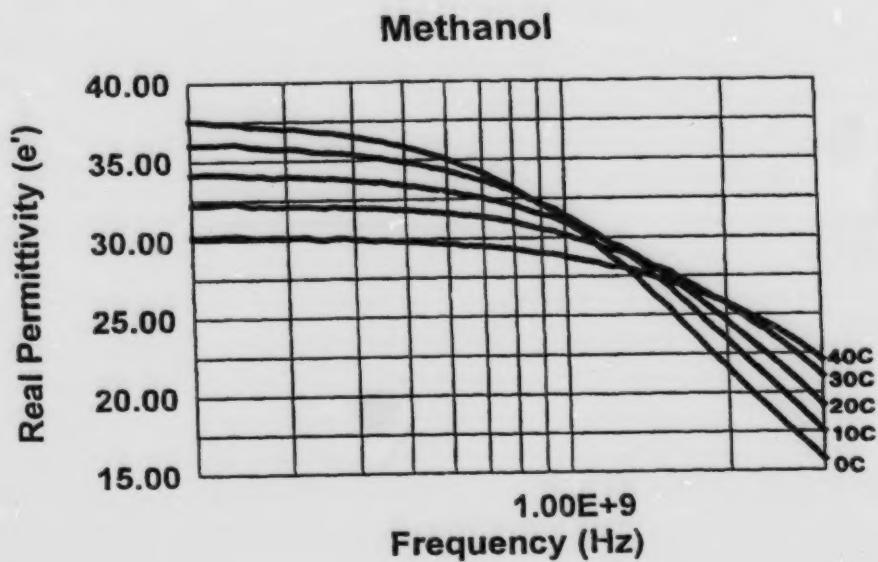
Project No. 972-1145
 Drawn GMC
 Reviewed 10/98
 Date



HP85070A DIELECTRIC PROBE

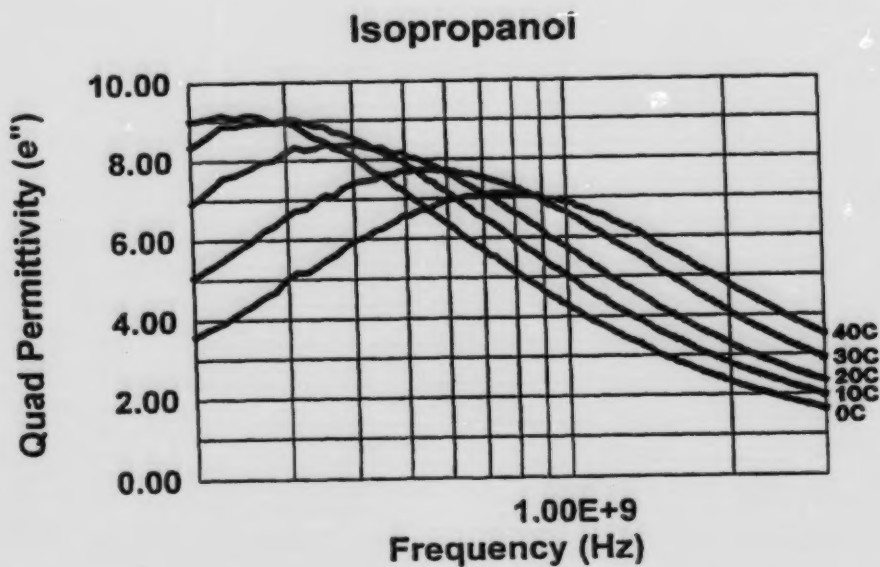
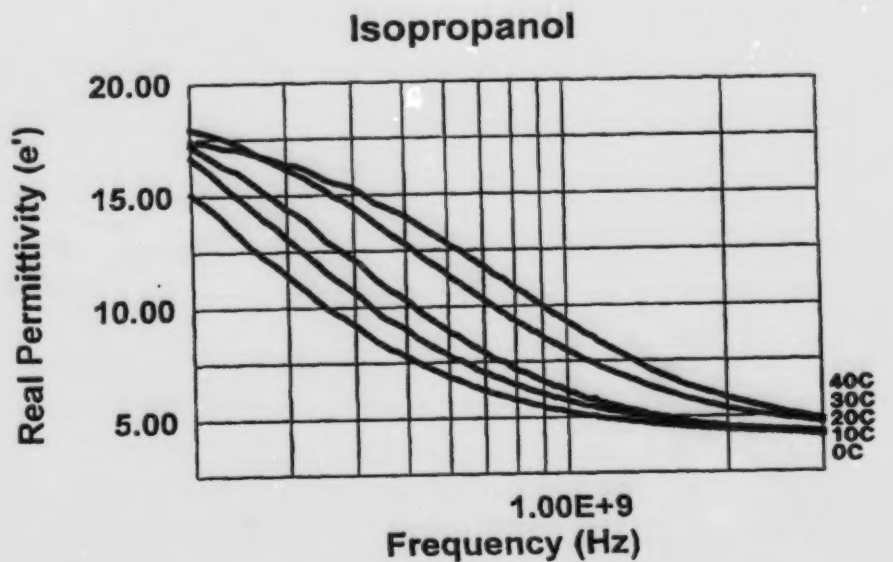


Project No. 072-1145 Drawn GMC Reviewed Date 5/98



**COMPLEX RELATIVE PERMITTIVITY
TEMPERATURE DEPENDENCE - METHANOL**

Project No. 878-1145 Drawn GMC Reviewed Date 5/93

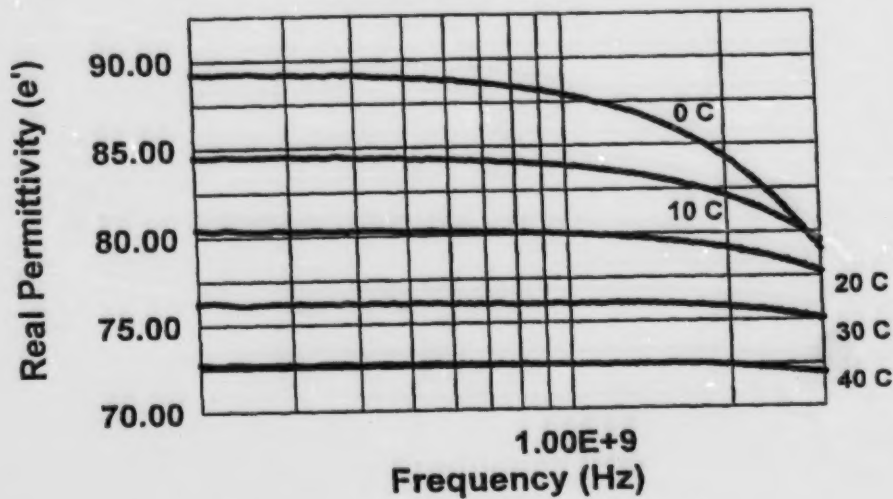


**COMPLEX RELATIVE PERMITTIVITY
TEMPERATURE DEPENDENCE - ISOPROPANOL**

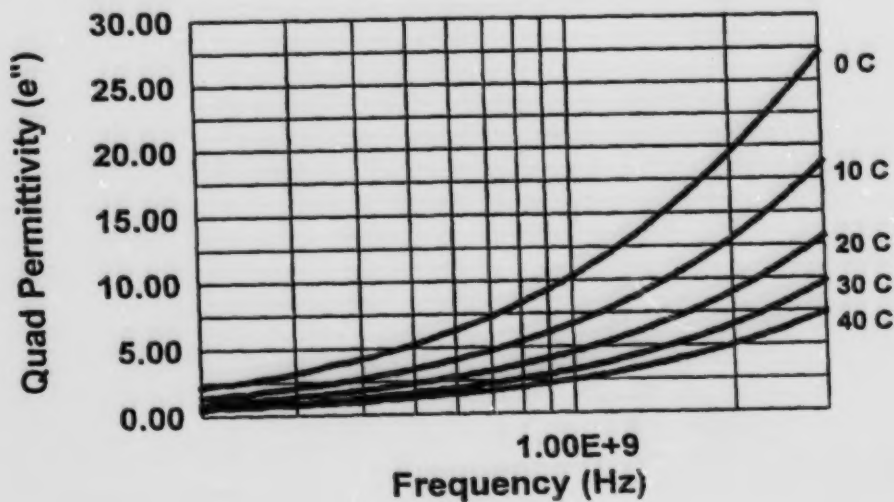
Figure

45

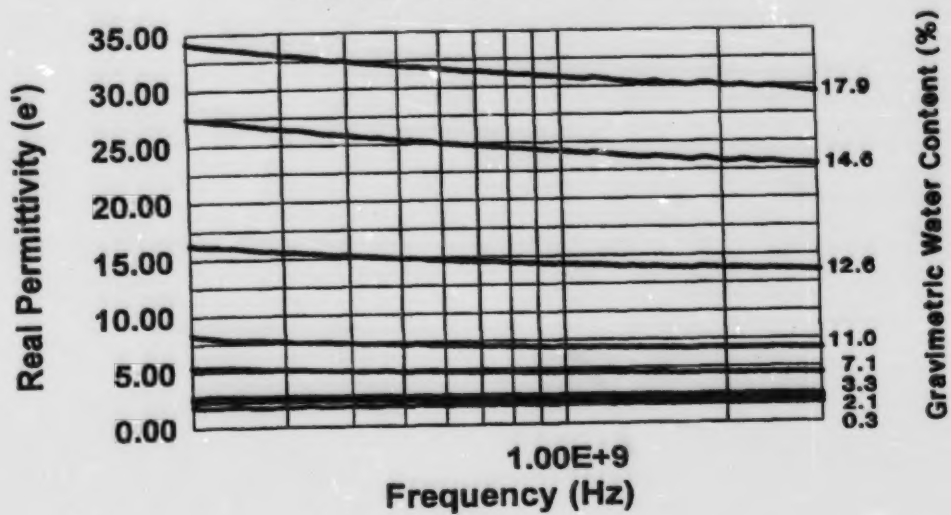
Distilled Water



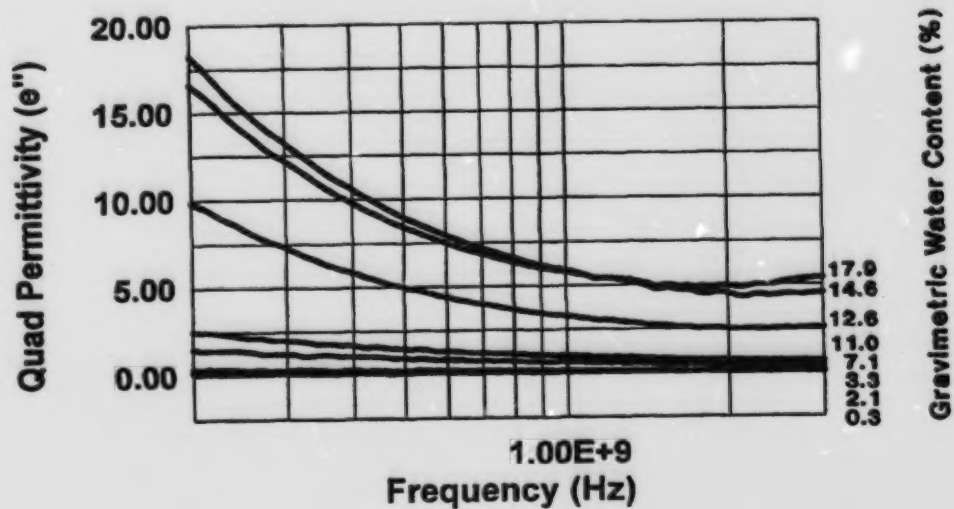
Distilled Water



Moisture Effect MLA-1

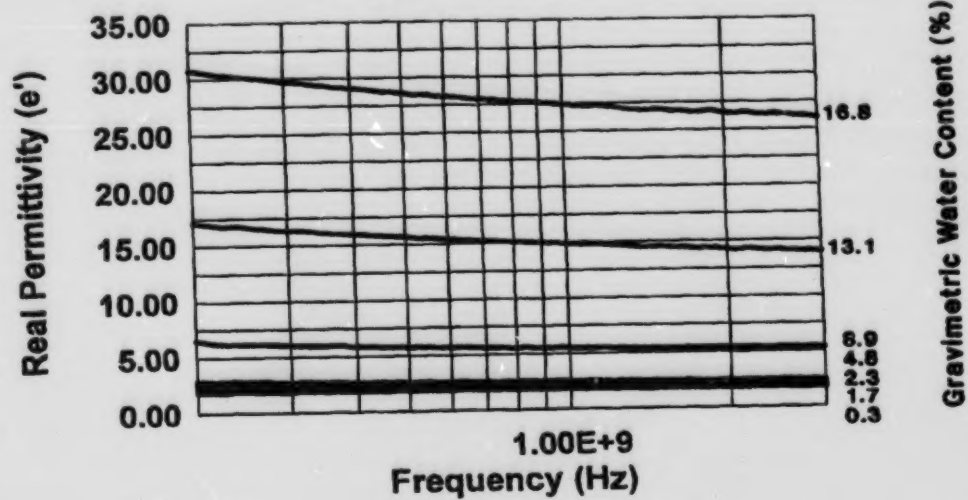


Moisture Effect MLA-1

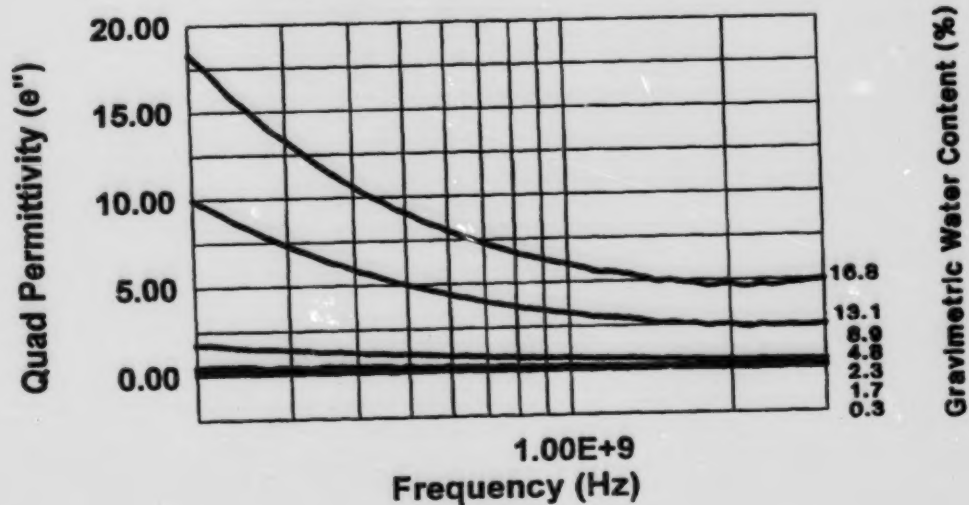


Project No. 872-1145 Drawn GMC Reviewed Date 20/7/98
 Project No. 872-1145 Drawn GMC Reviewed Date 20/7/98

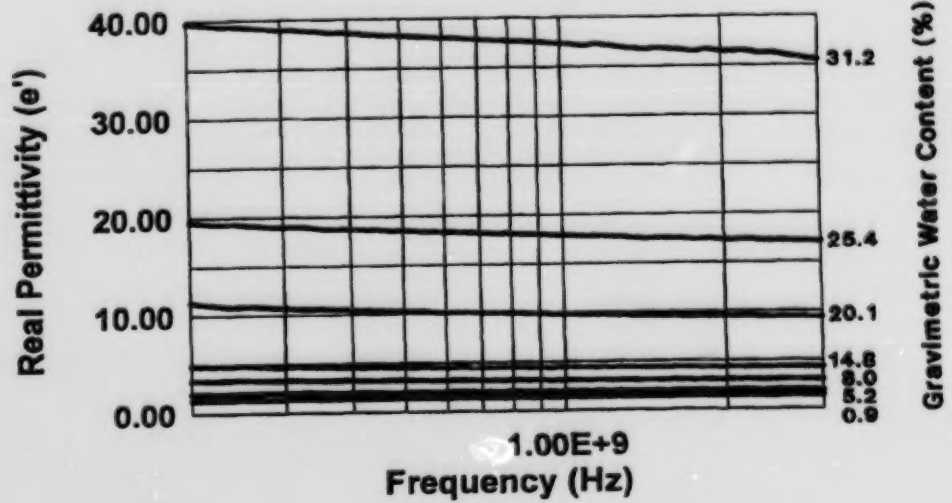
Moisture Effect MLA-2



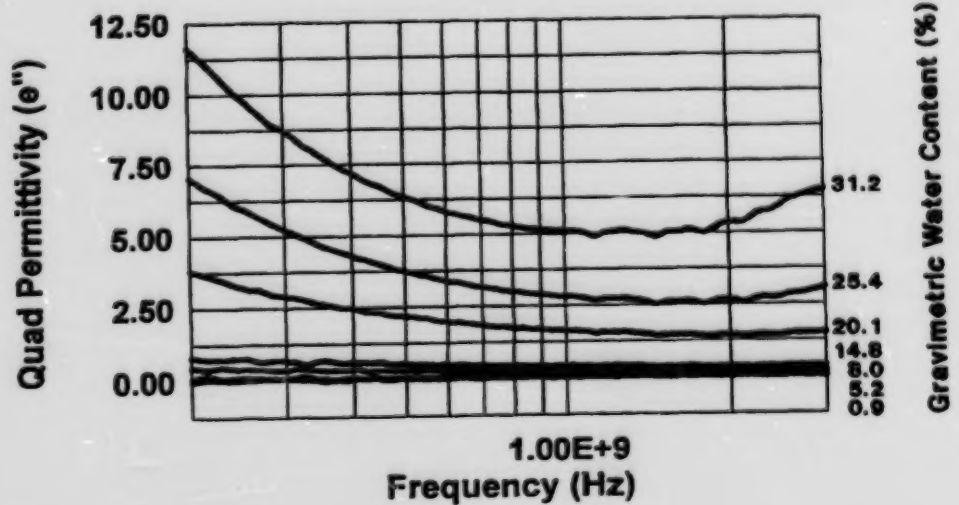
Moisture Effect MLA-2



Moisture Effect NS-1



Moisture Effect NS-1



Project No. 872-1145 Drawn GAC Reviewed Date 20/7/98

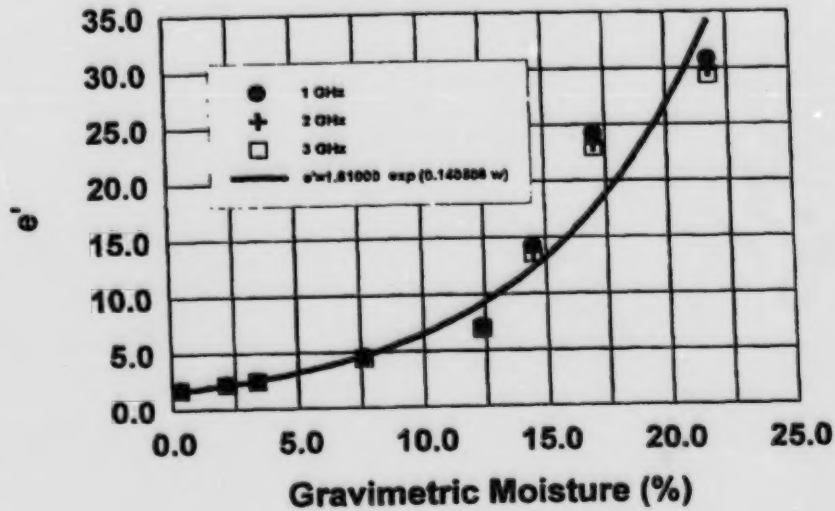


COMPLEX PERMITTIVITY - NS-1

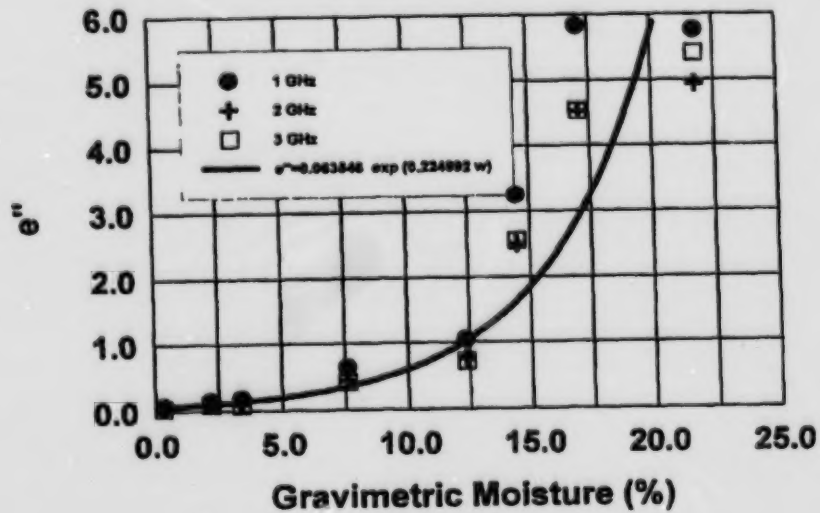
Figure 49

Project No. _____ Date _____
 Drawn _____ Reviewed _____

MLA-1

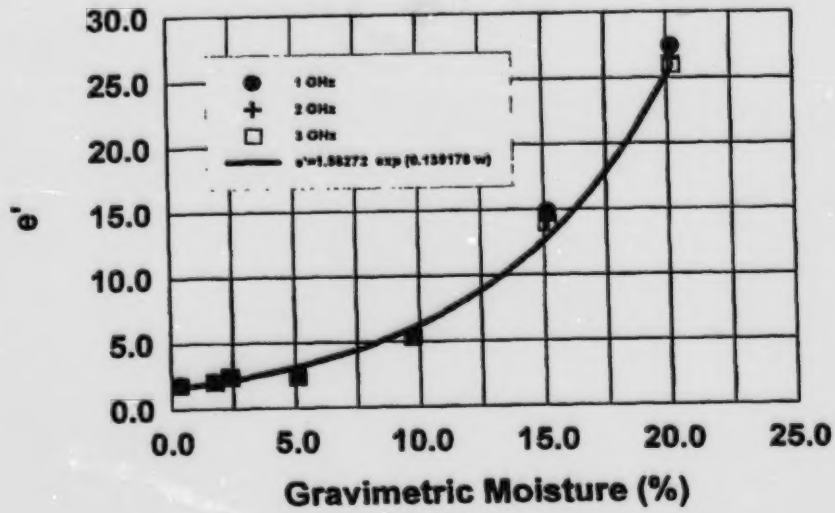


MLA-1

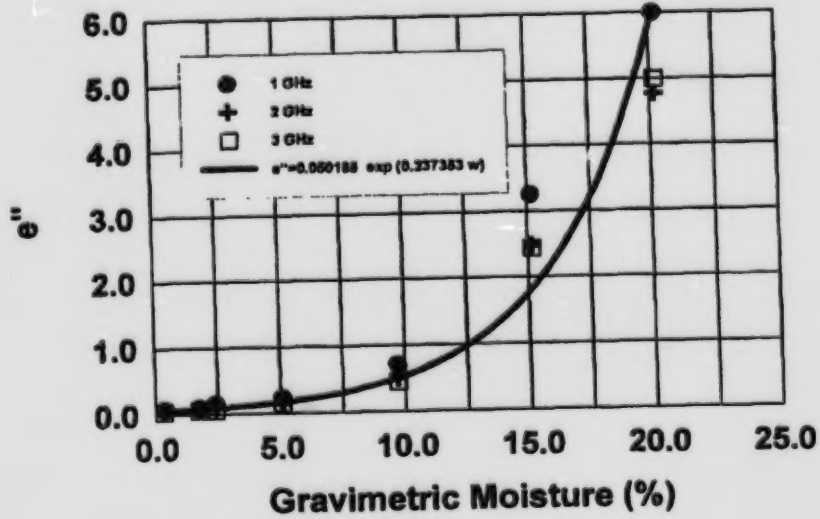


Project No. _____ Drawn _____ Reviewed _____ Date _____

MLA-2

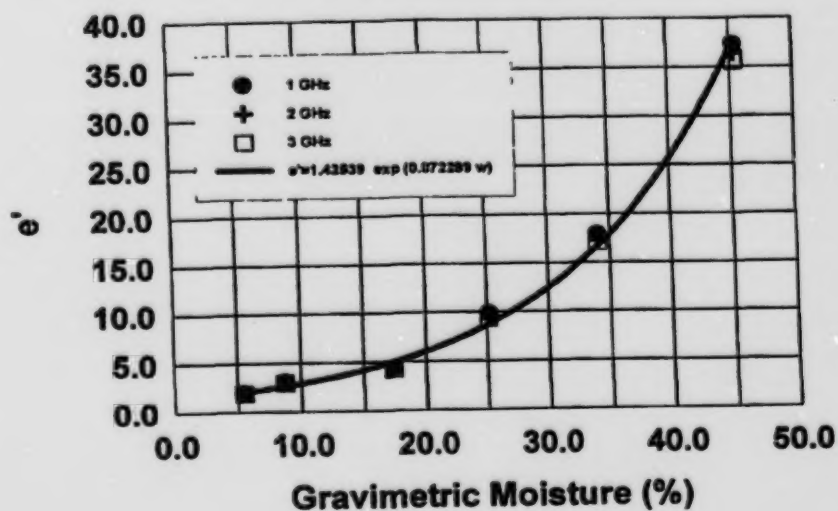


MLA-2

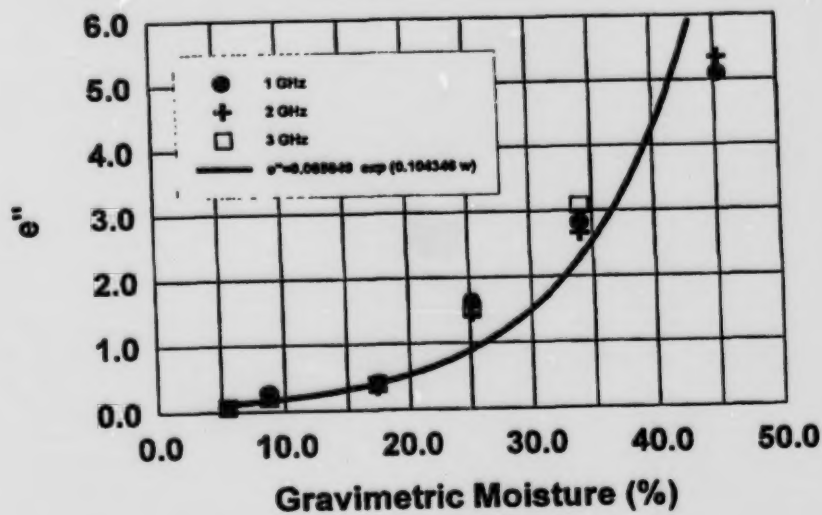


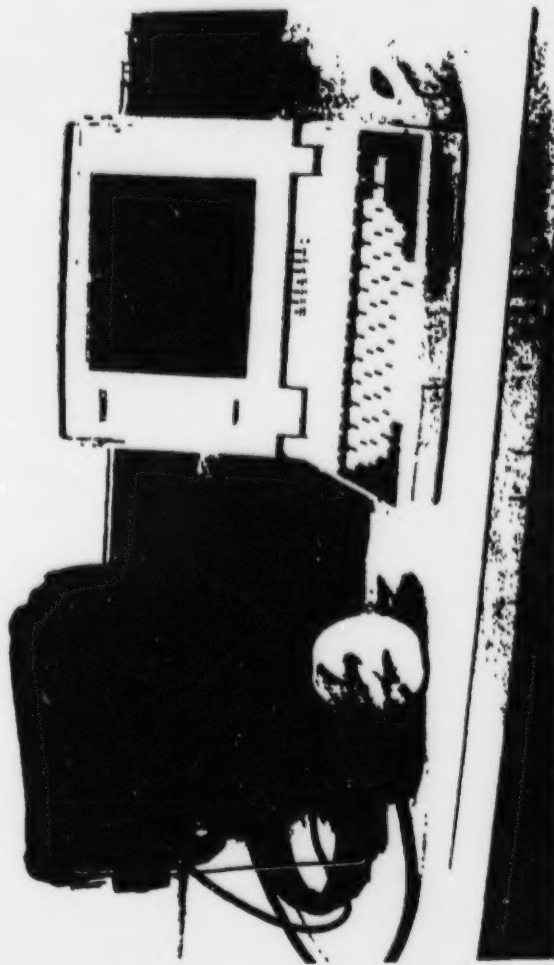
Project No. _____ Drawn _____ Reviewed _____ Date _____

NS-1



NS-1

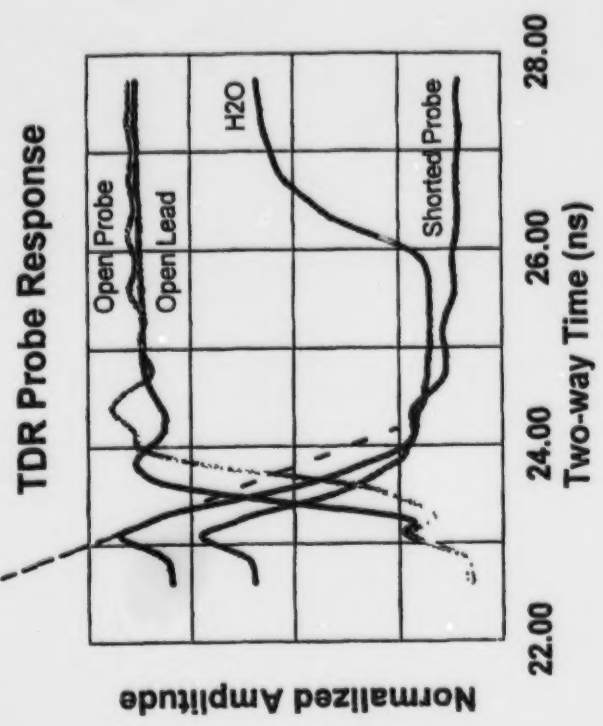




Project No. 972-1145
Drawn GMC
Reviewed



TIME DOMAIN REFLECTOMETRY (TDR) APPARATUS

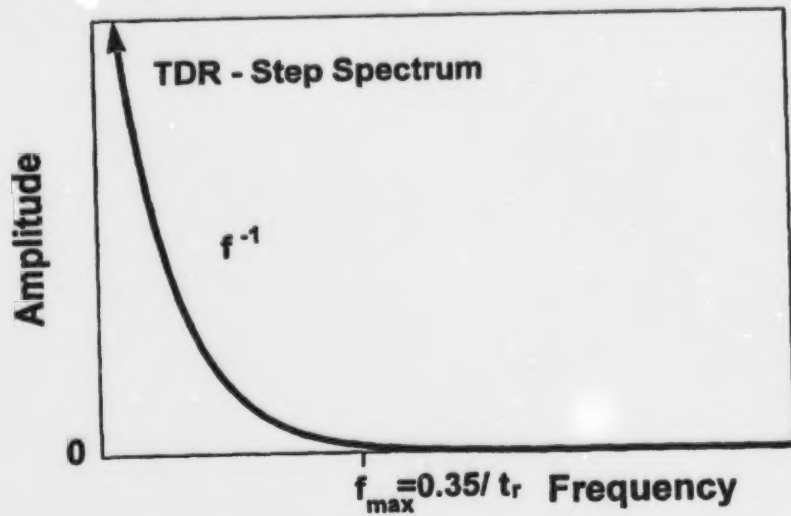
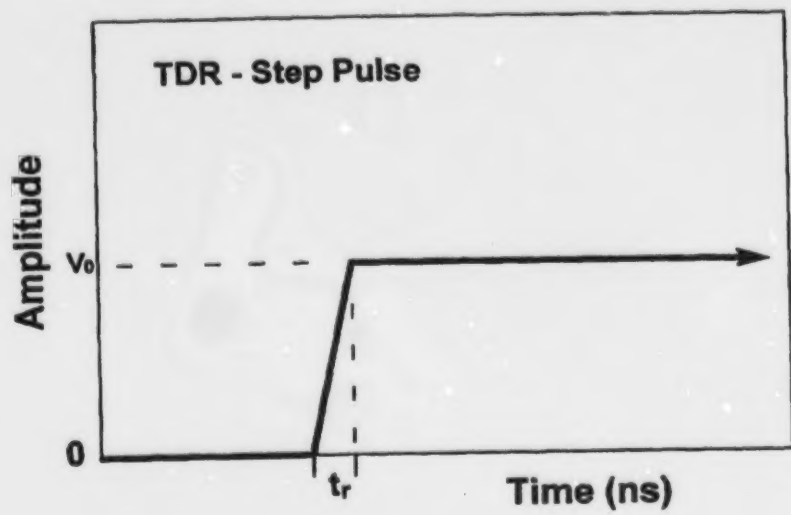


Project No. 073-1145
 Drawn GWC
 Reviewed *[Signature]*
 04/10/00



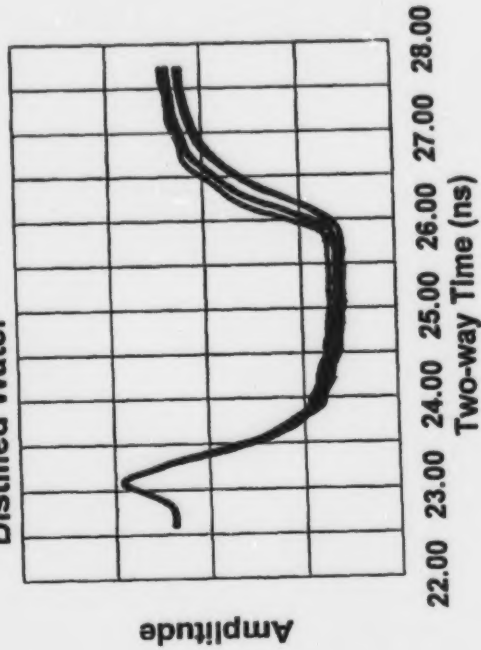
TDR TEST PROBE-CELL

Project No. Drawn Reviewed Date

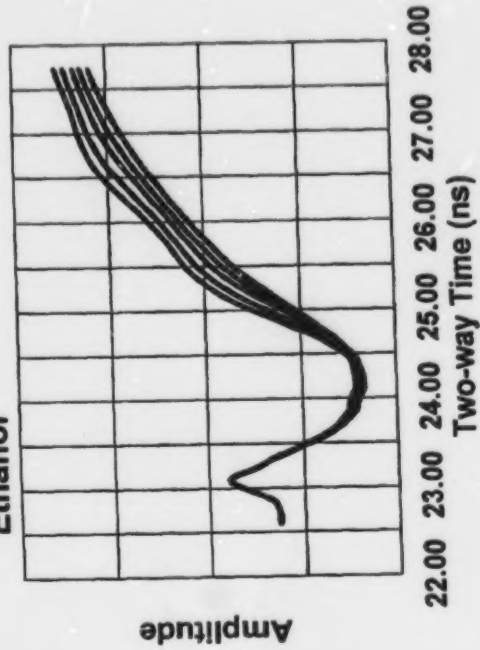


(Tektronix, 1999)

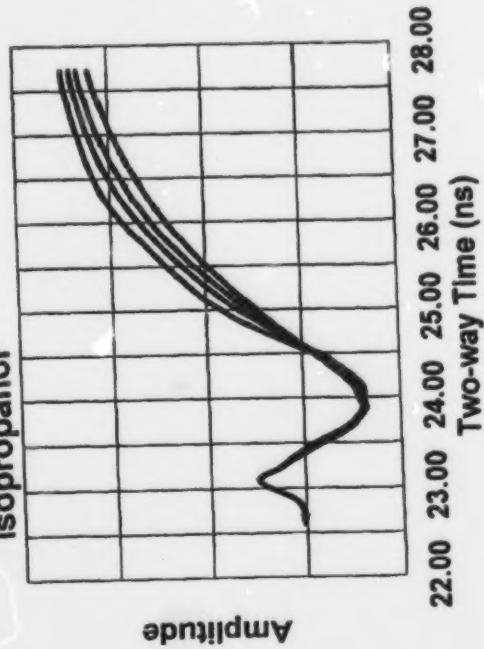
Distilled Water



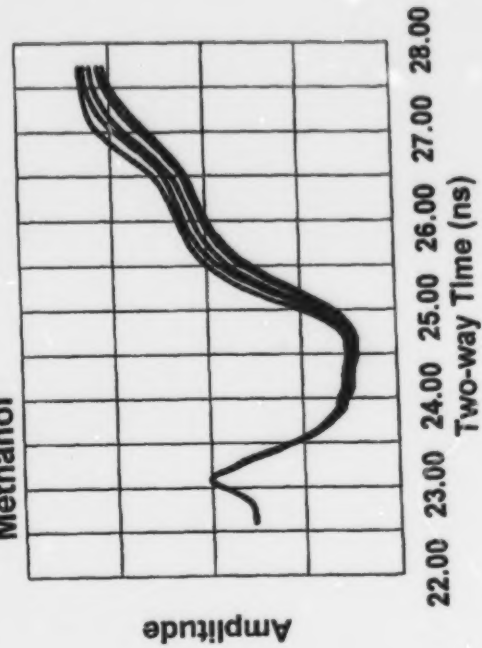
Ethanol



Isopropanol



Methanol



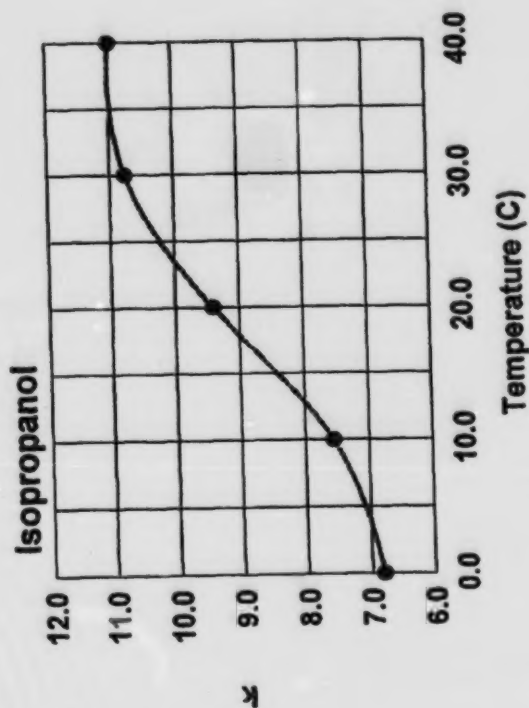
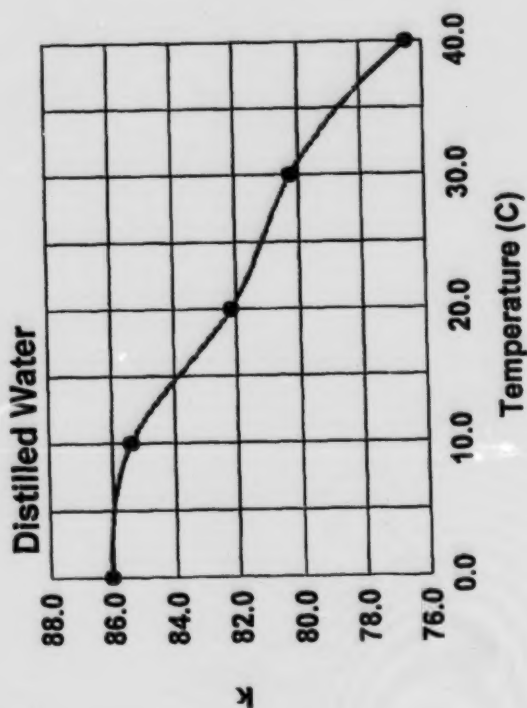
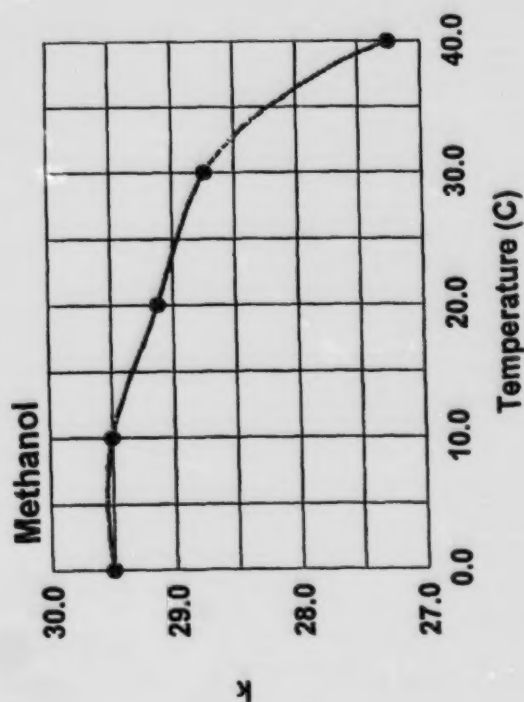
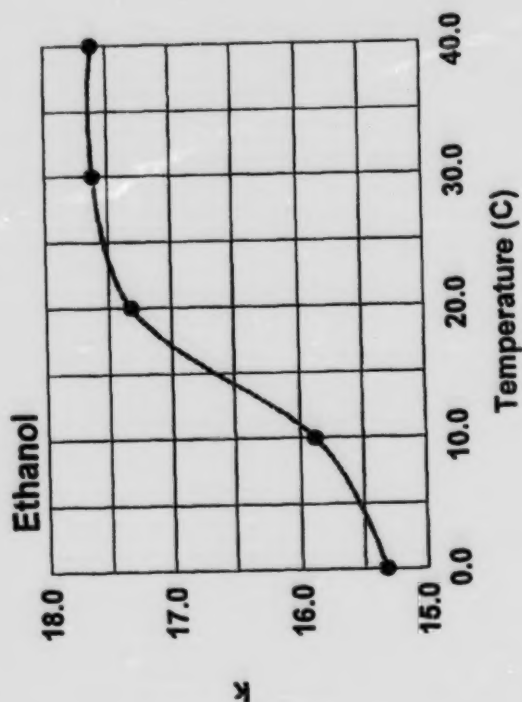


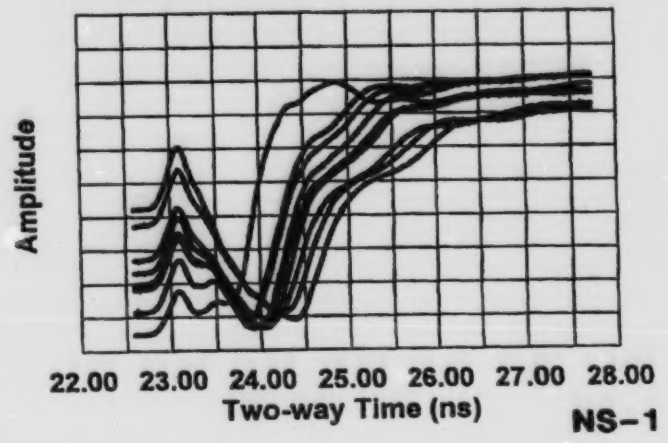
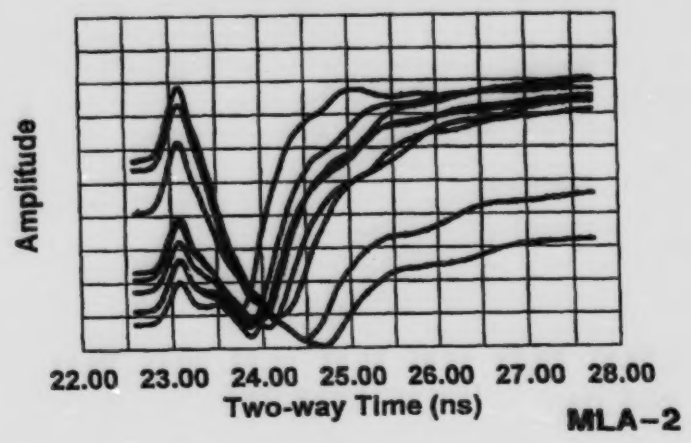
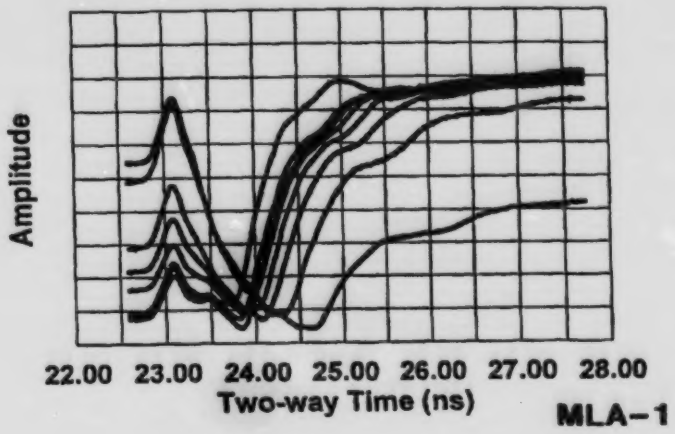
Figure 56

TDR DIELECTRIC CONSTANT - TEMPERATURE

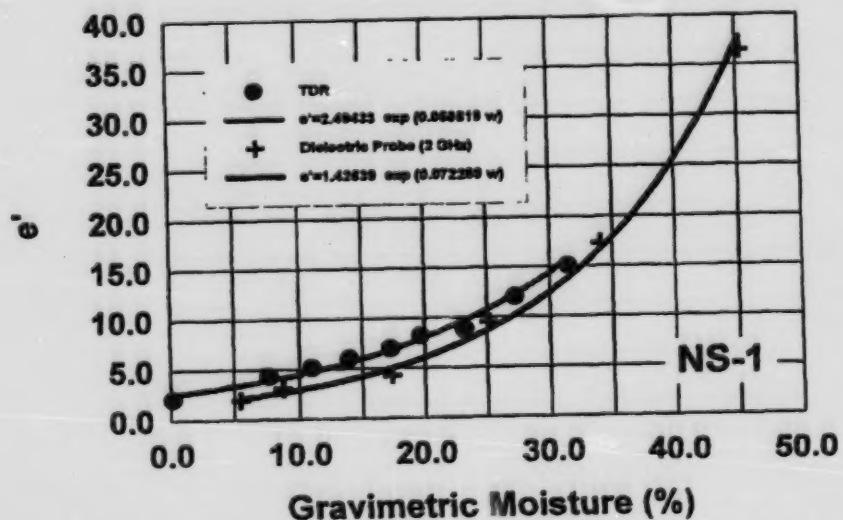
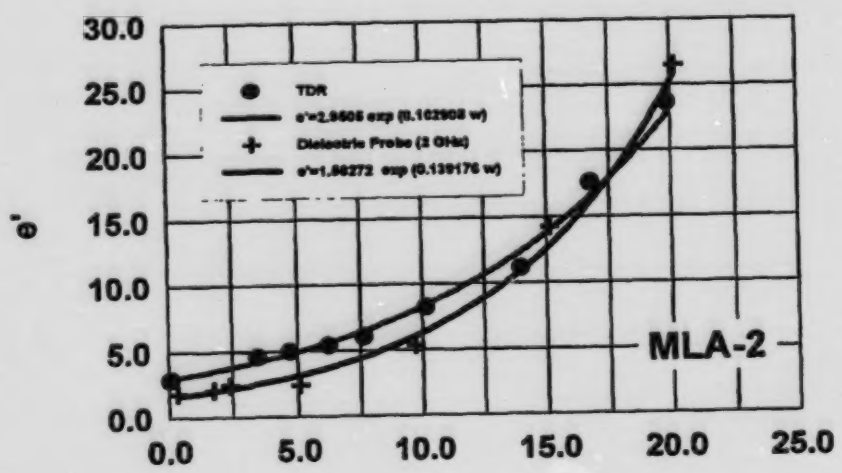
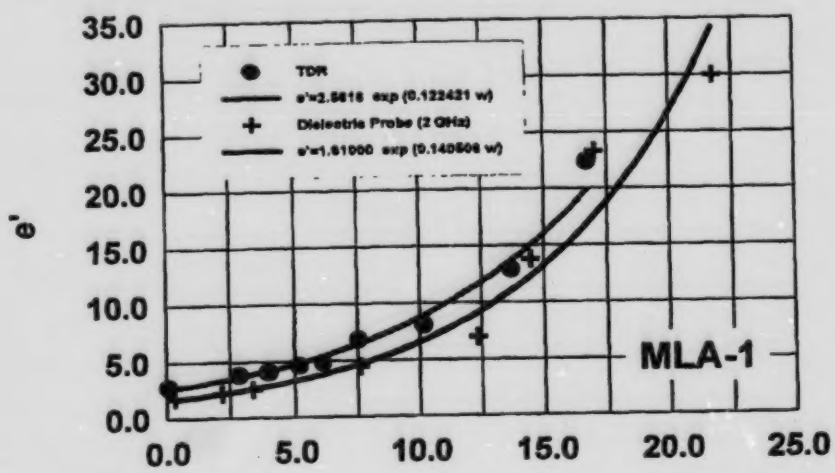


Project No.
 Drawn
 Reviewed

Project No. 972-1145 Drawn *CMC* Reviewed *CK* Date 10/98

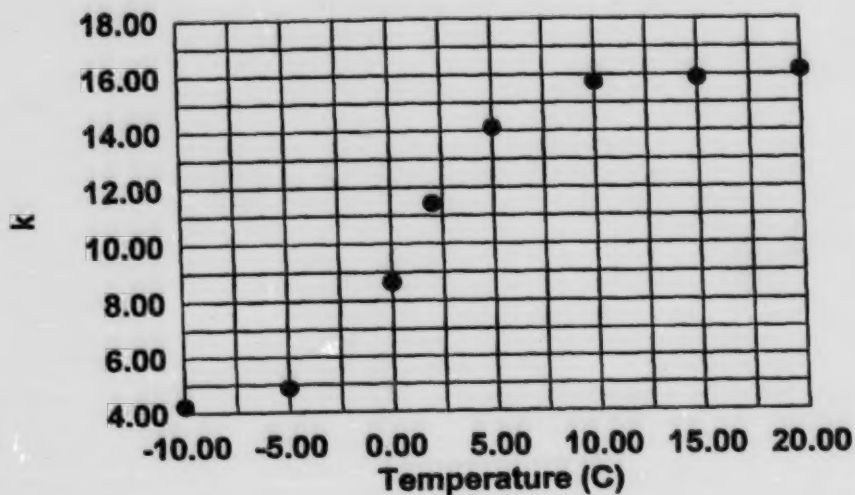


Project No. _____ Drawn _____ Reviewed _____ Date _____

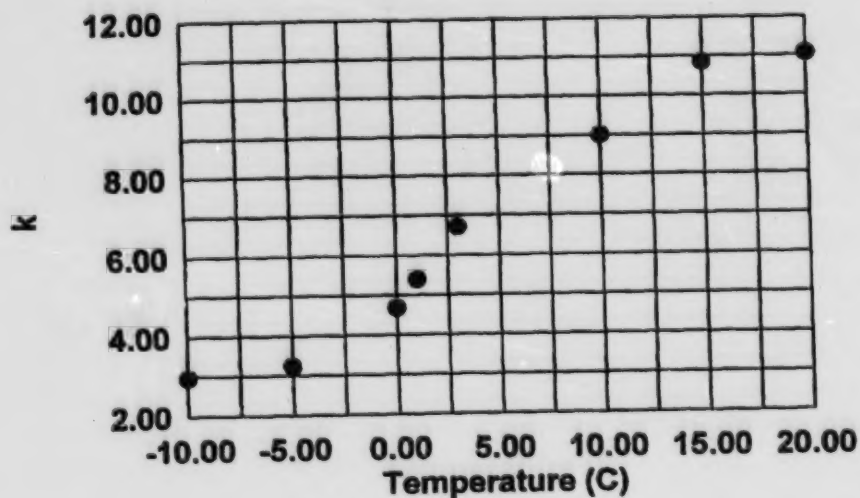


**TDR DIELECTRIC CONSTANT
GRAVIMETRIC MOISTURE**

MLA-1



NS-1

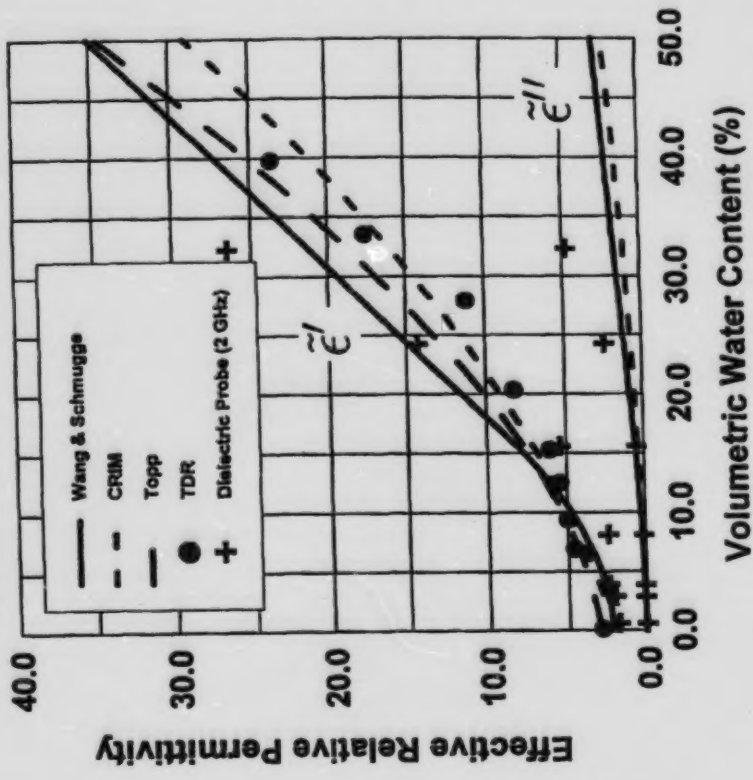


MLA-1 Sand / Silt / Clay = 0.9 / 0.08 / 0.02

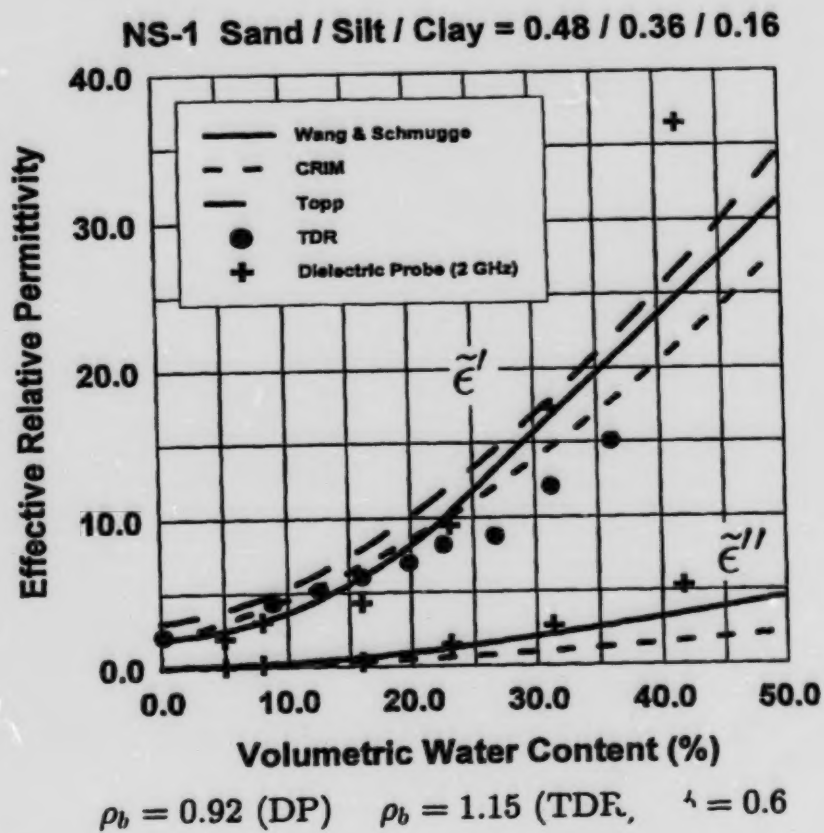


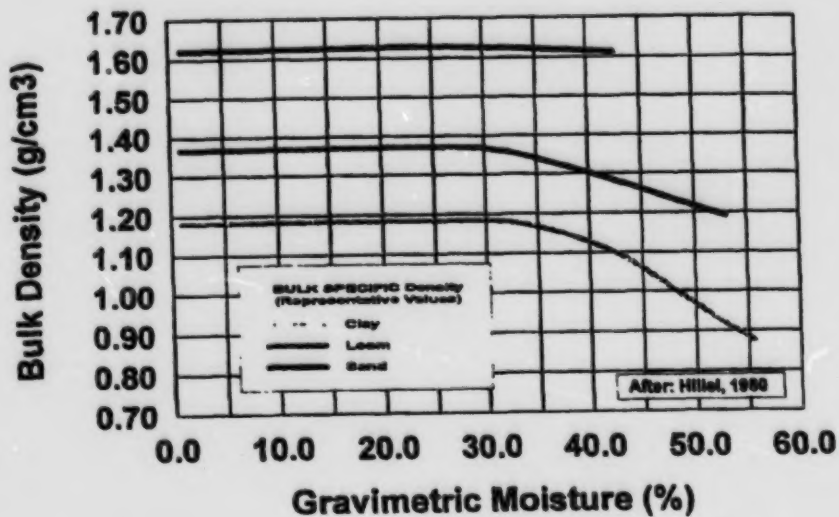
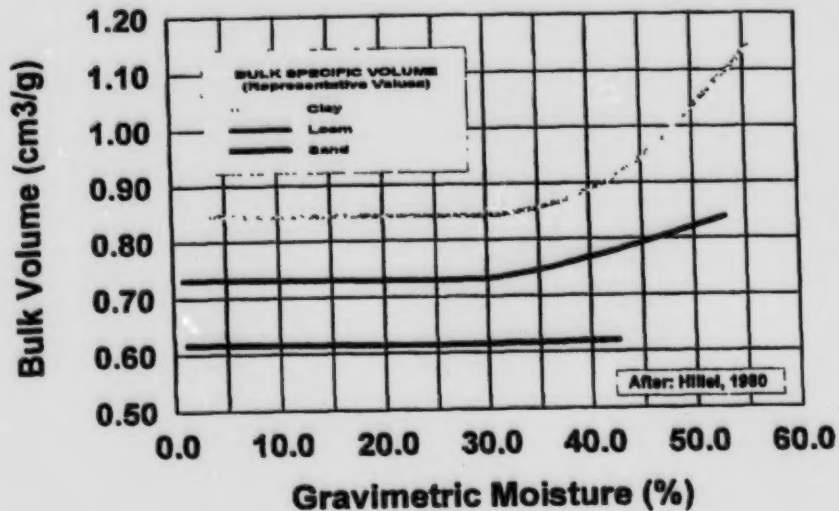
$\rho_b = 1.73$ (DP) $\rho_b = 1.99$ (TDR) $\phi = 0.5$

MLA-2 Sand / Silt / Clay = 0.9 / 0.1 / 0.0

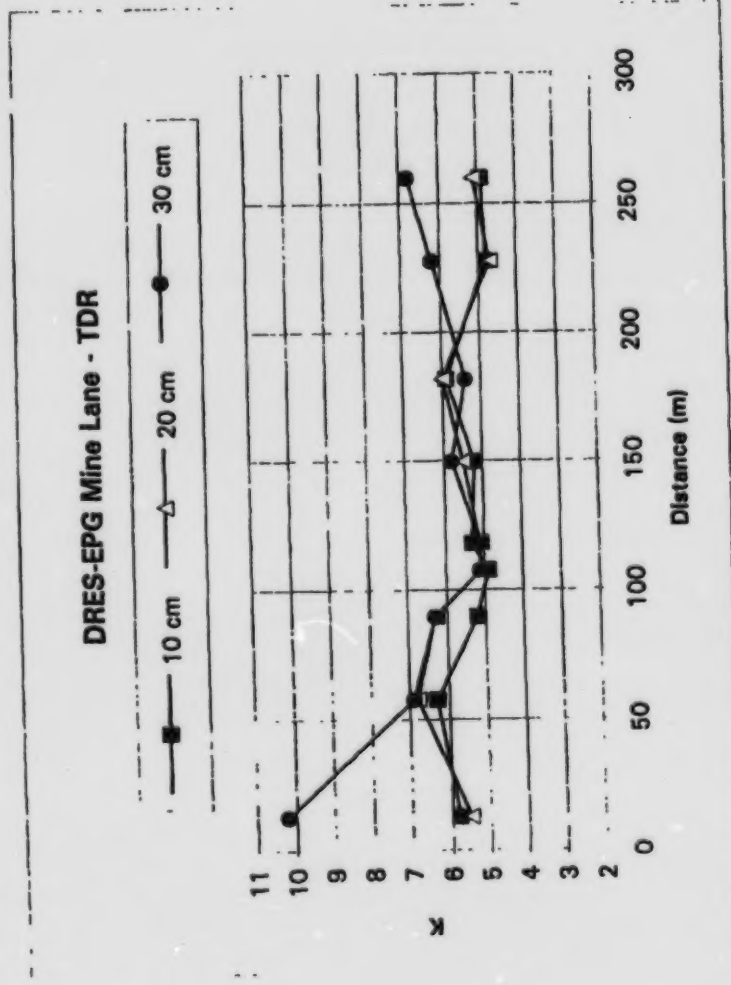


$\rho_b = 1.60$ (DP) $\rho_b = 2.00$ (TDR) $\phi = 0.5$

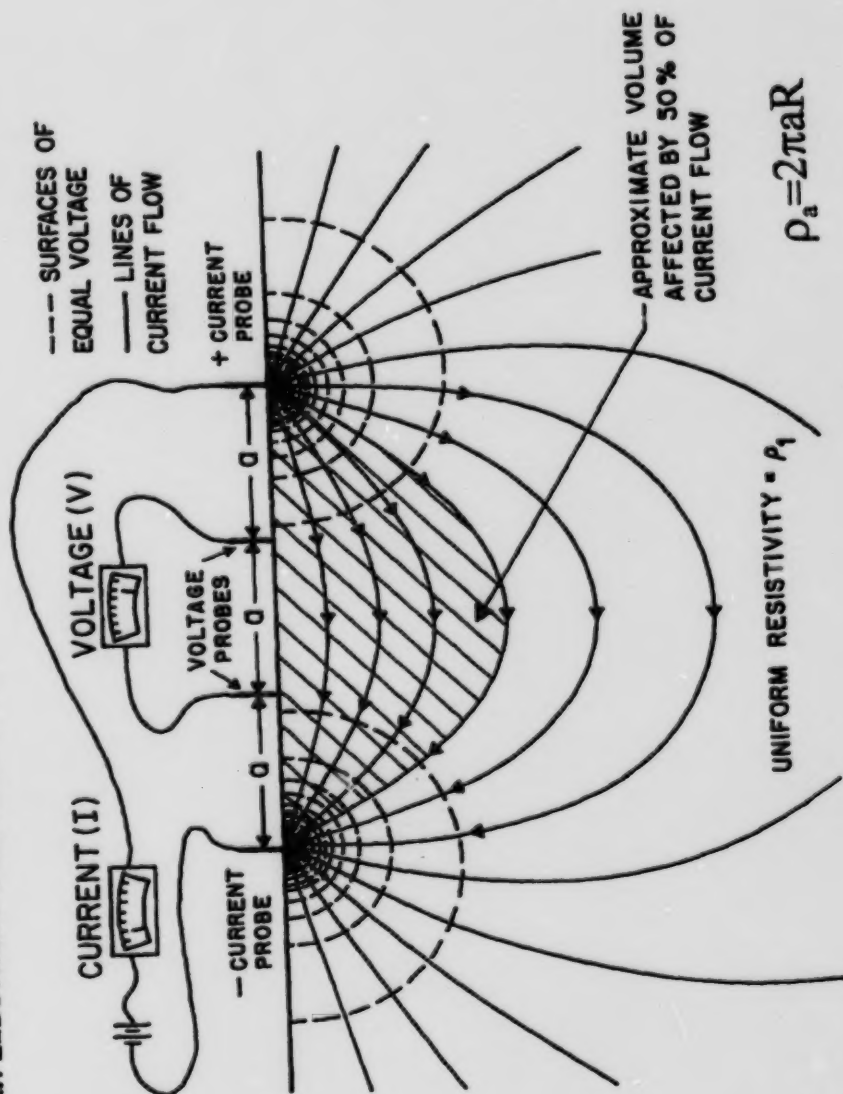




NOTE: Data acquired 26-28/8/98

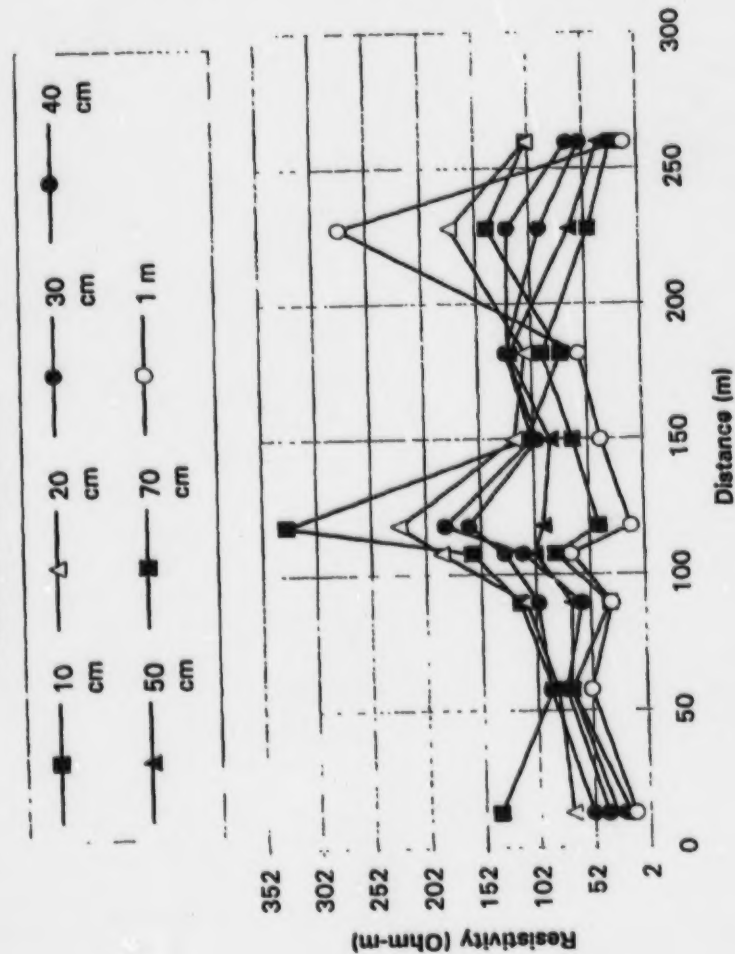


WENNER ELECTRODE CONFIGURATION



NOTE: Data acquired 26-28/8/98

DRES-EPG Mine Lane - Resistivity

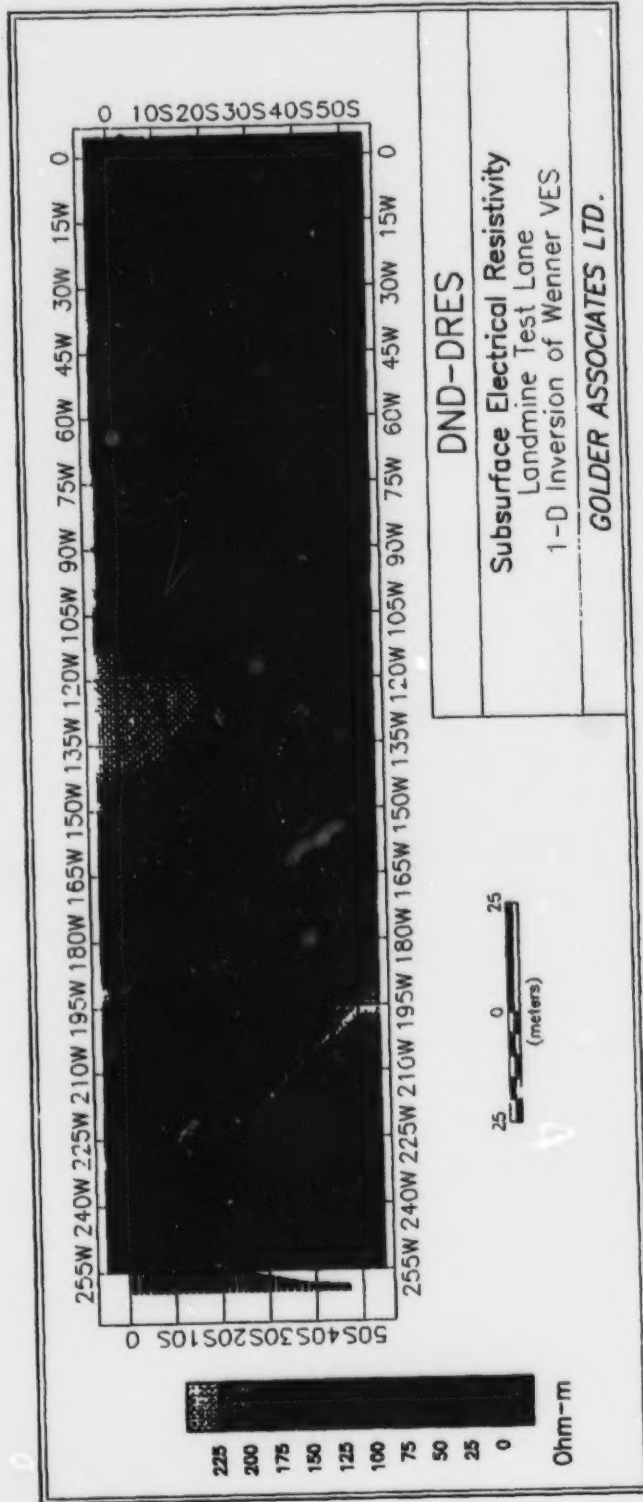


Project No. 973-1145
 Drawn GMC
 Reviewed



DRES MINE LANE - RESISTIVITY PROFILES

Figure



NOTE: Contours and colour levels are approximate and based on interpolation - polynomial regression from depicted 1-D electrical resistivity - depth profiles. Uncertainty increases with distance from constraints.



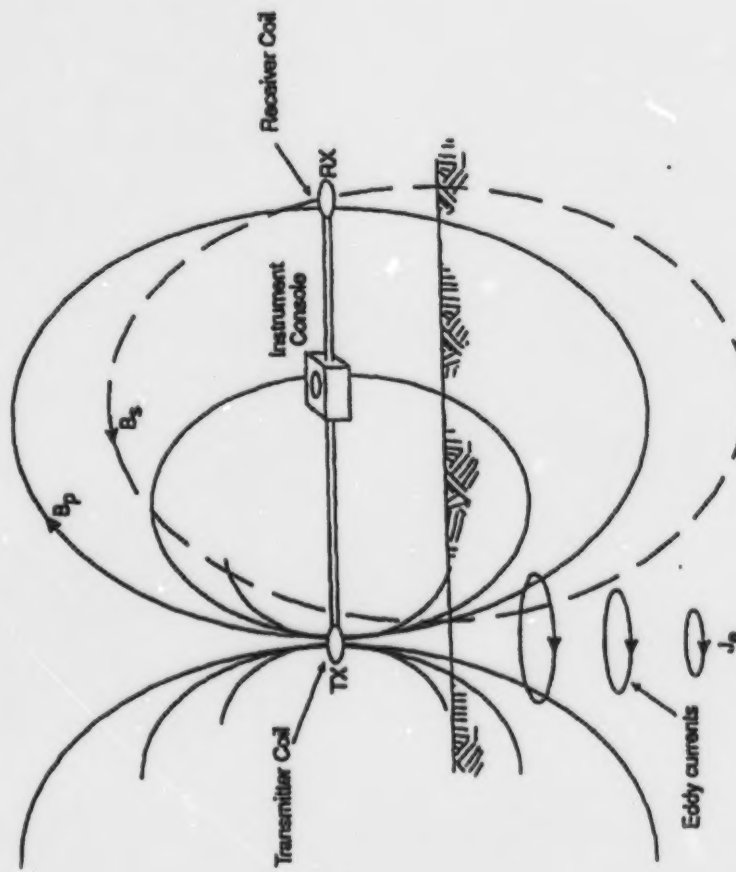
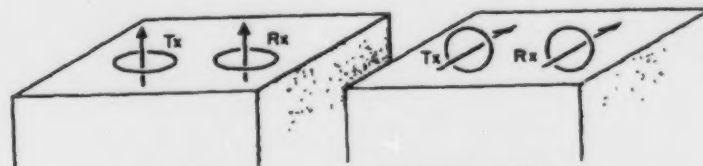


Figure 68

ELECTROMAGNETIC MEASUREMENT



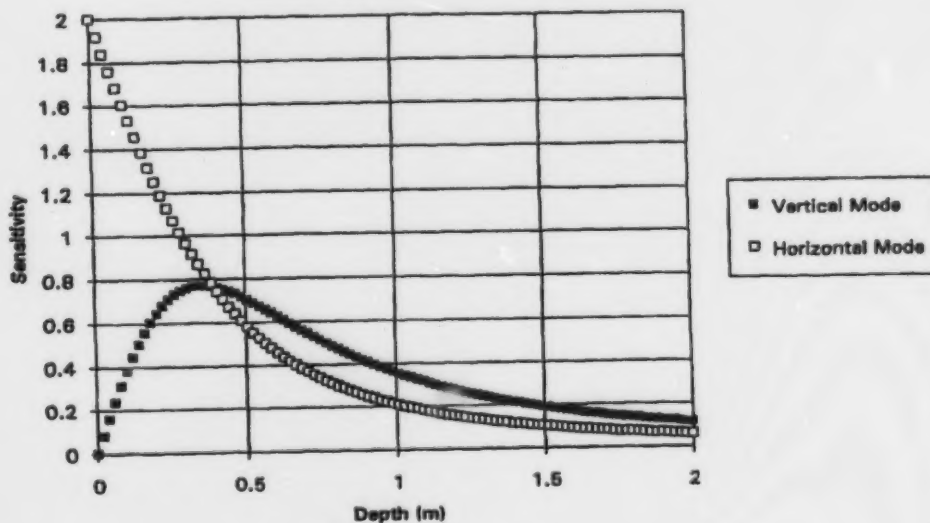
Project No. 842-1848
 Drawn GMC
 Reviewed 1/11/01



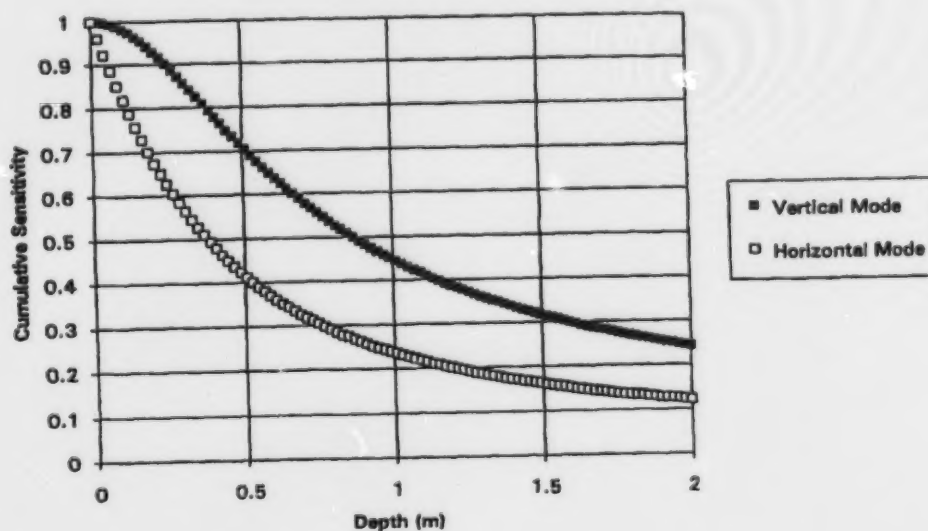
VERTICAL DIPOLE

HORIZONTAL DIPOLE

EM-38 Depth Sensitivity



EM-38 Cumulative Sensitivity



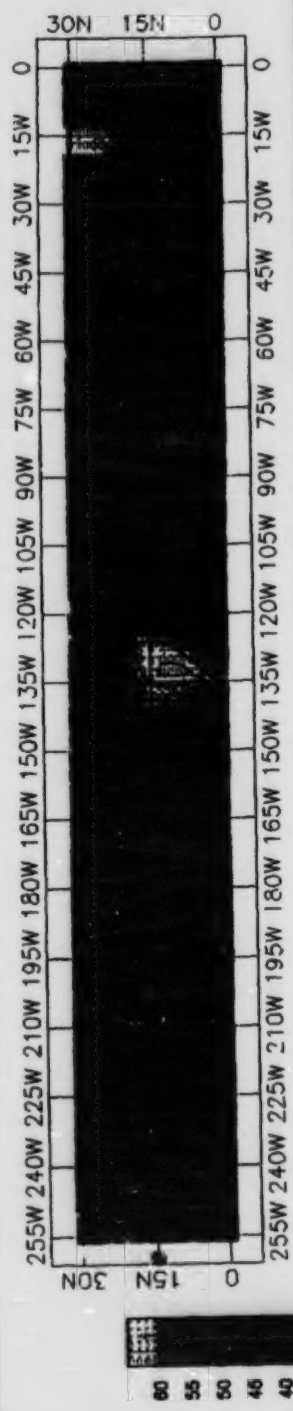
Project No. 042-1848 Drawn GMC Reviewed Date 1/8/04



EM-38 RESPONSE CHARACTERISTICS

Figure

69

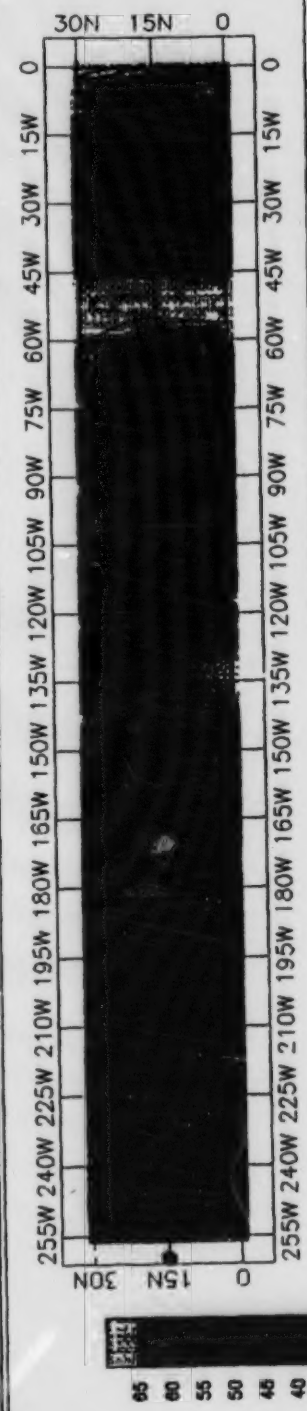


DND - DRES

EM-38 Apparent Soil Conductivity (mS/m)
Landmine Test Lane
Horizontal Dipole Mode

GOLDER ASSOCIATES LTD.

NOTE: ● Denotes TDR/VES Test Location.



DND - DRES

EM-38 Apparent Soil Conductivity (mS/m)
Landmine Test Lane
Vertical Dipole Mode

GOLDER ASSOCIATES LTD.

NOTE: ● Denotes TDR/VES Test Location.



VTS - EXCAVATED TEST BED

Figure

71



Project No. _____
Drawn by _____
Reviewed by _____
Date _____

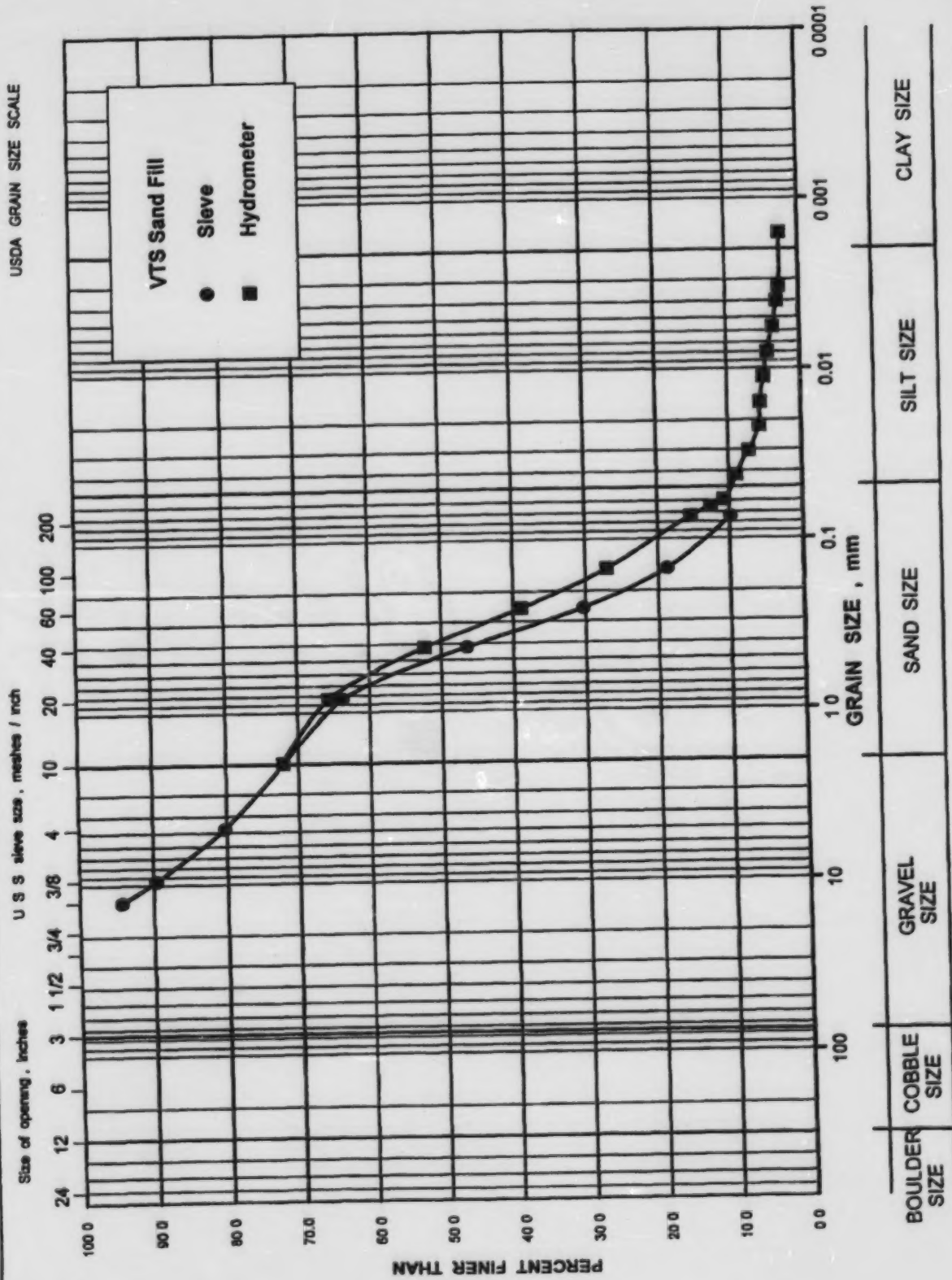


Figure 72

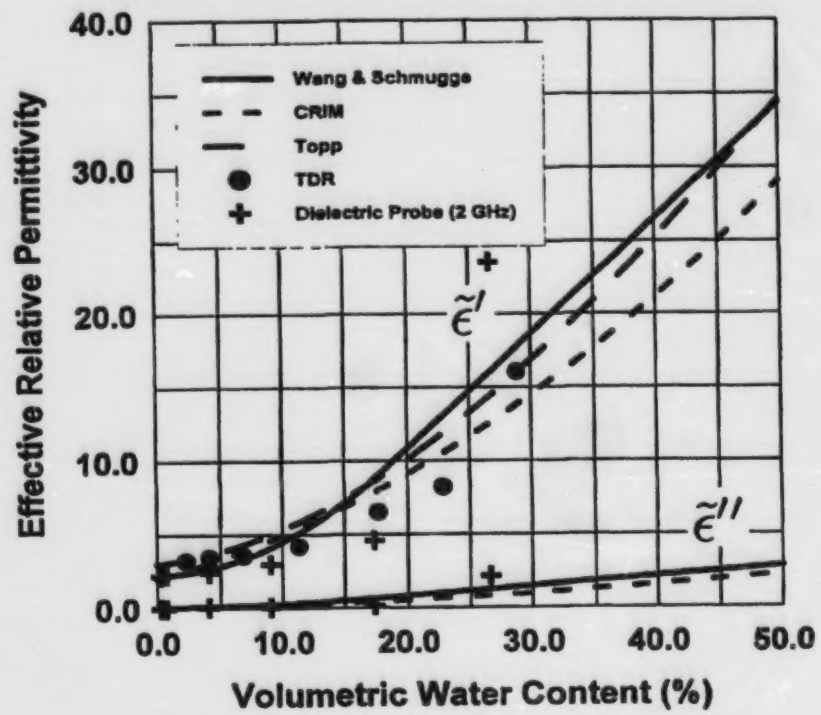
GRAIN SIZE DISTRIBUTION VTS SAND FILL



Project No. 9721145
 Drawn JG
 Reviewed WMG
 06/17/00

Project No. Drawn Reviewed Date

VTS-1 Sand / Silt / Clay = 0.9 / 0.07 / 0.03

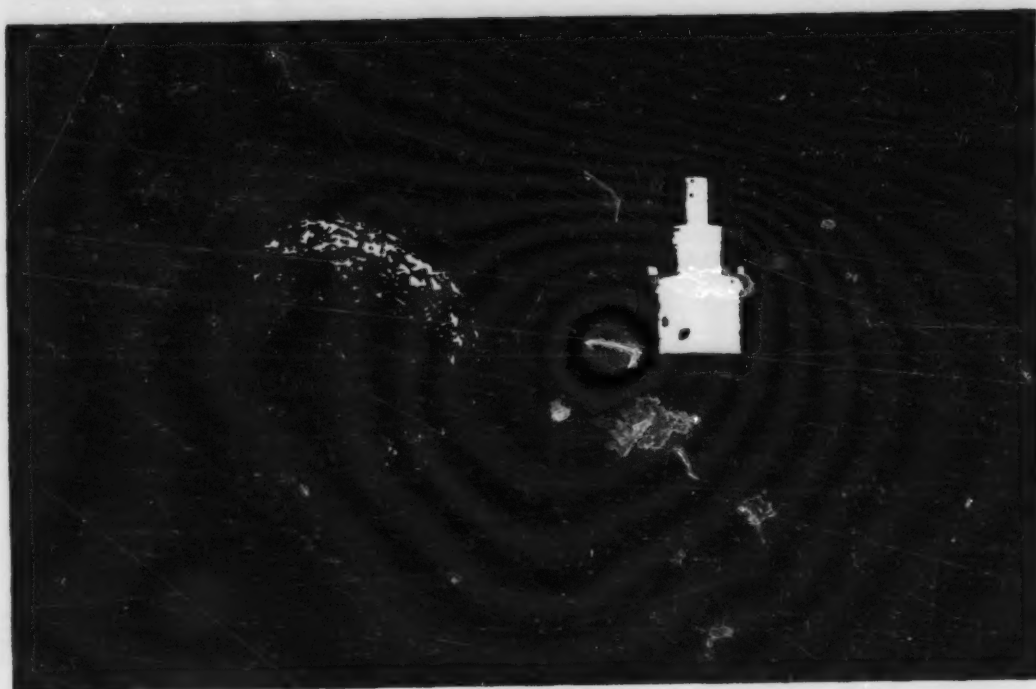
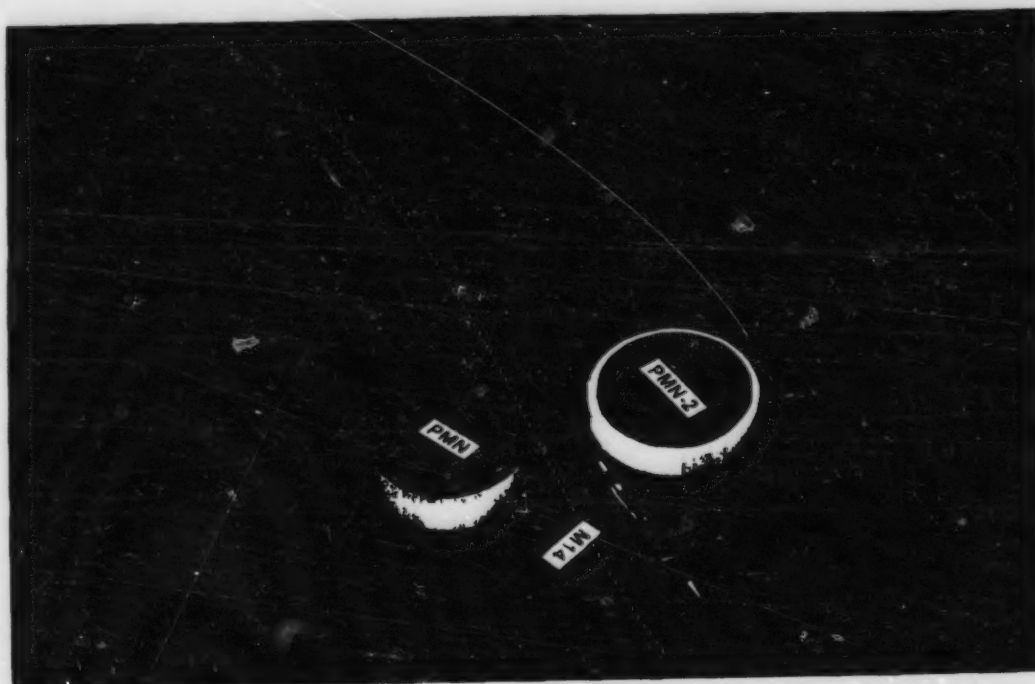


$\rho_b = 1.73$ (DP) $\rho_b = 2.14$ (TDR) $\phi = 0.5$



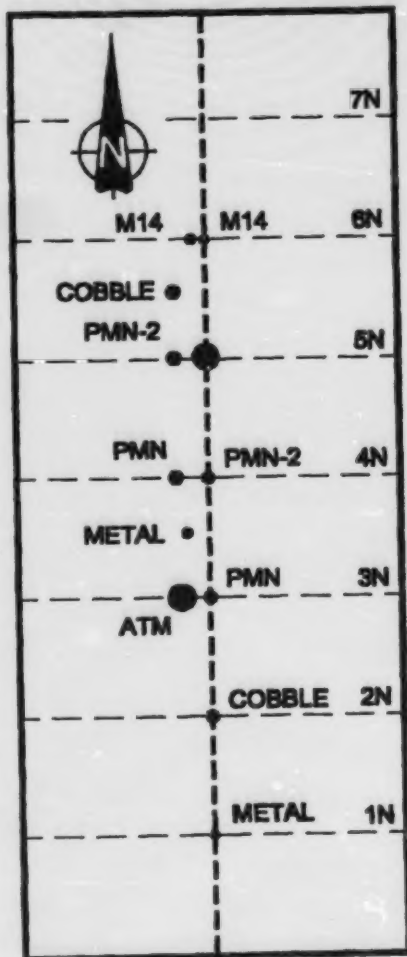
PERMITTIVITY - MOISTURE
VTS SAND FILL

Figure
73



Project No. 872-1149
 LATENT
 A.S.
 NOVEMBER 2003
 G.S.L.
 L.S.B.
 September 1999

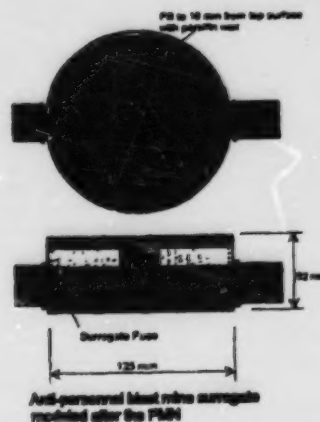
VANCOUVER TEST SITE (VTS)



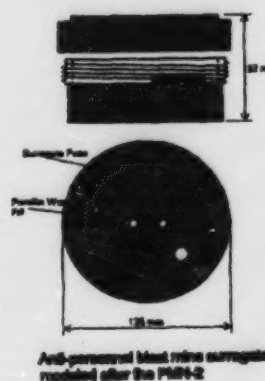
VTS-1 ●
 VTS-2 ●

Note: All surrogates 10 cm depth.

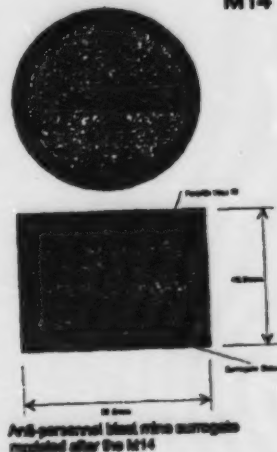
PMN

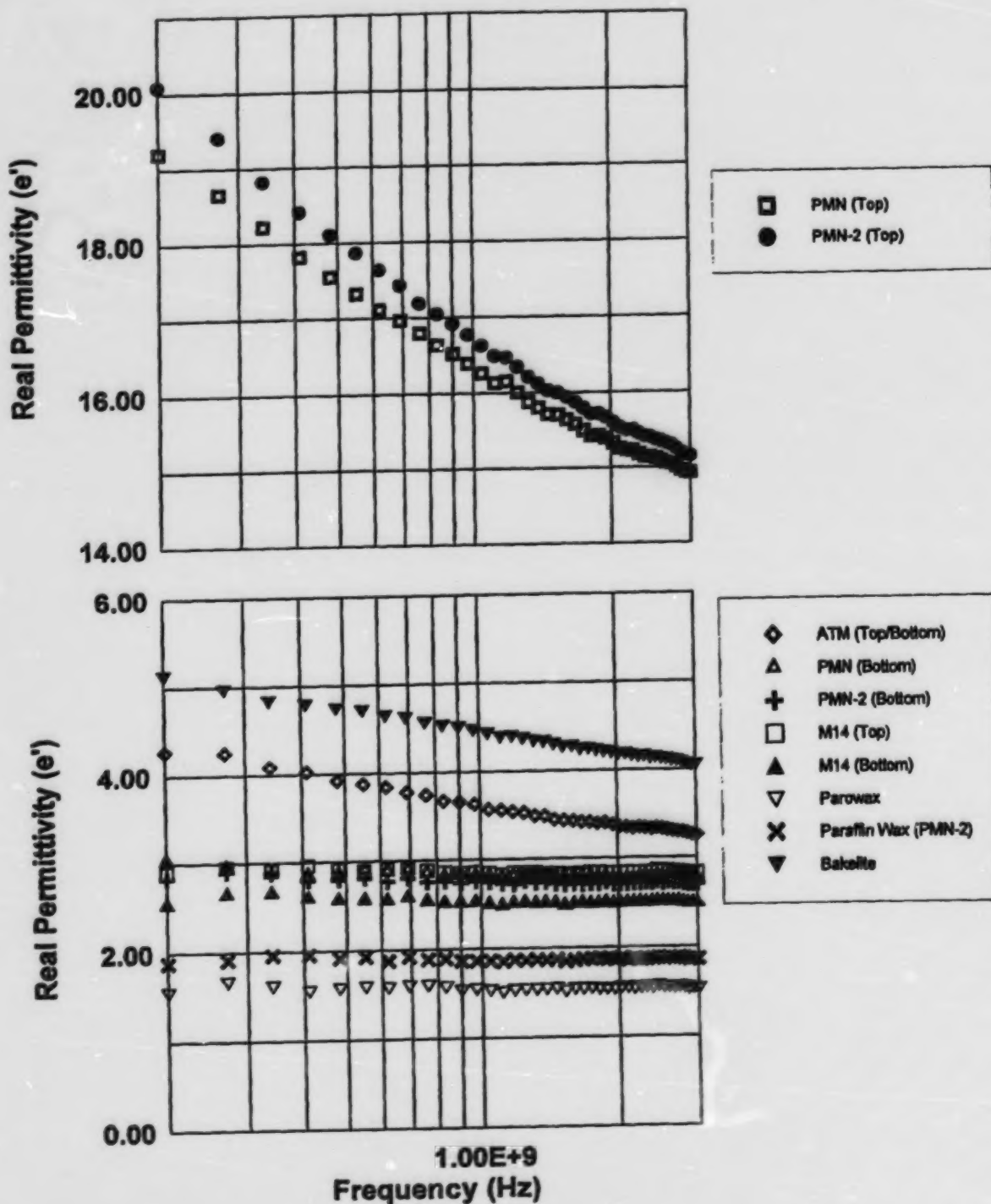


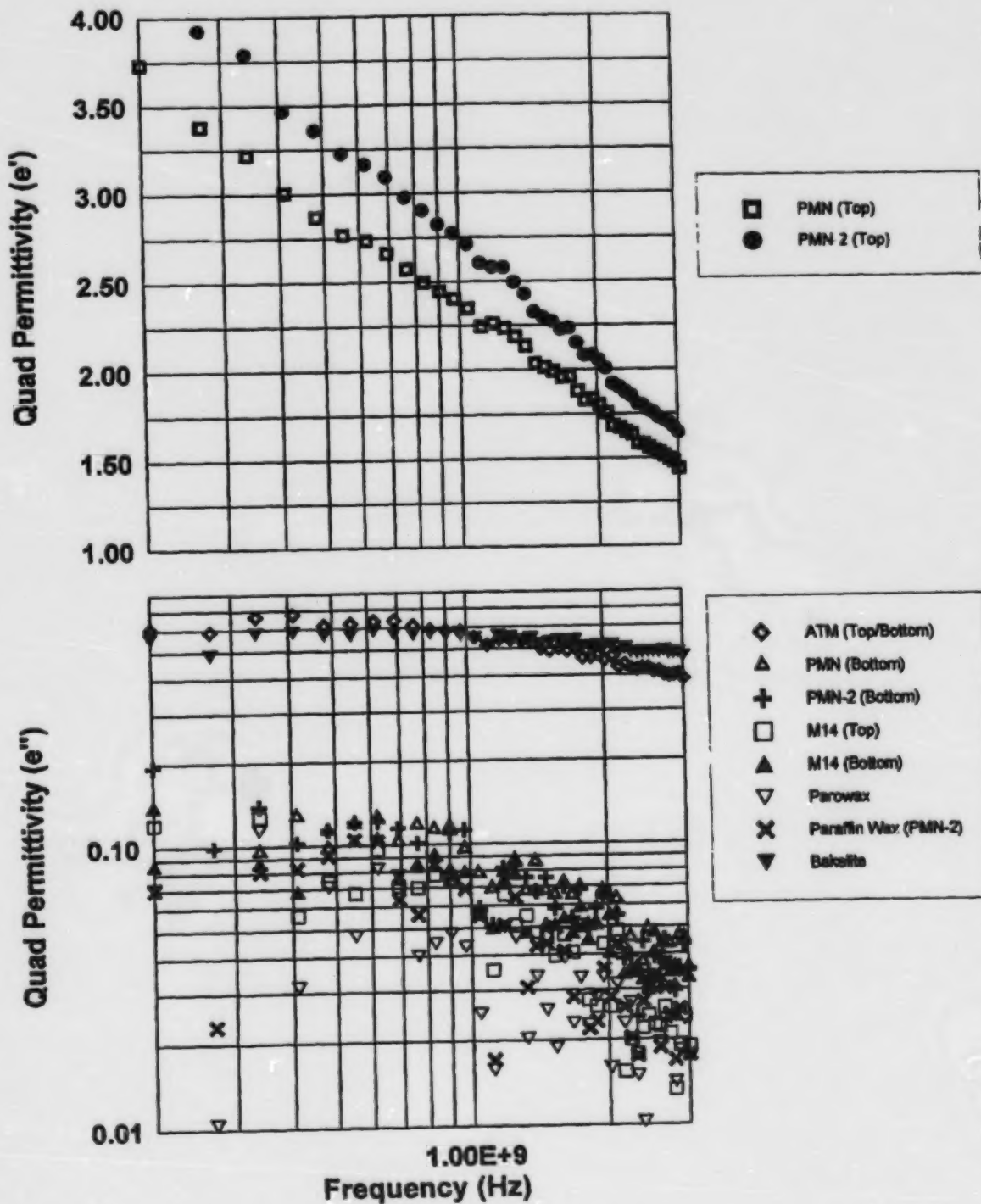
PMN-2



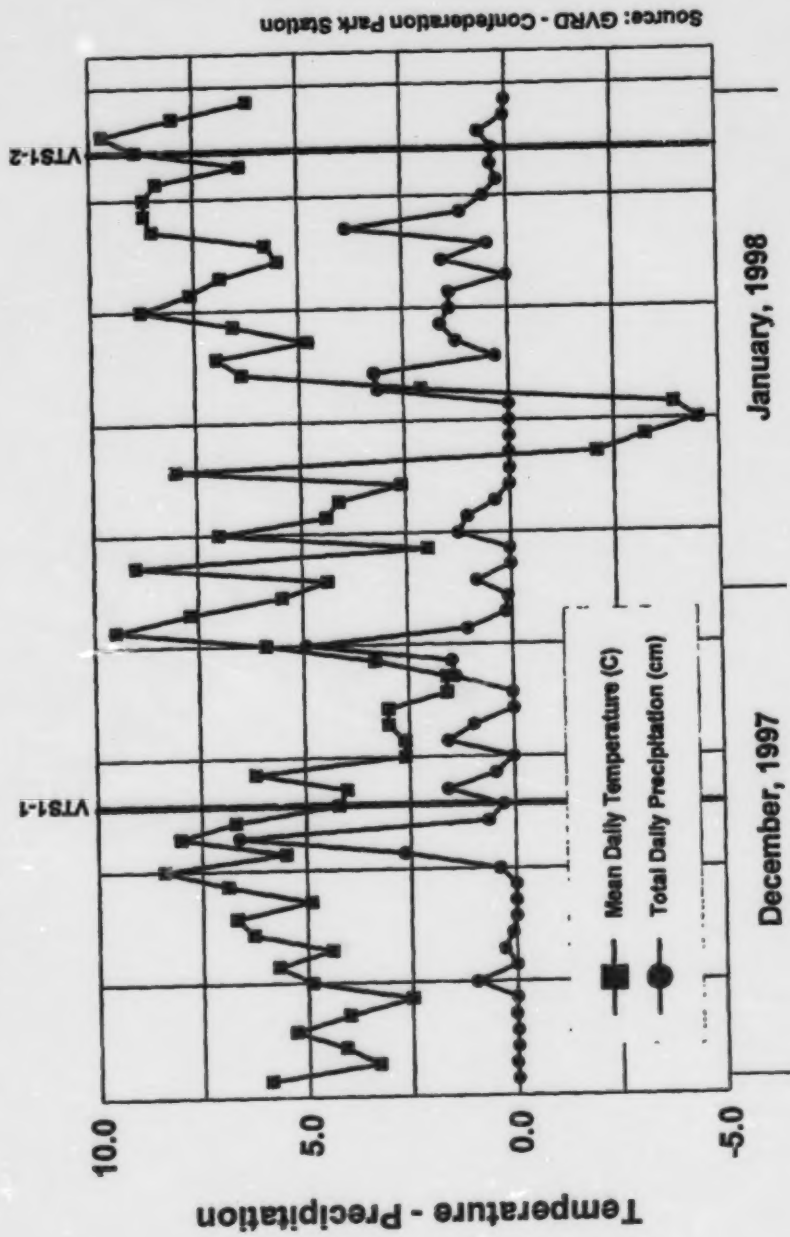
M14







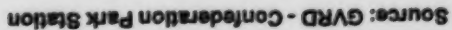
VTS1 Environmental Data



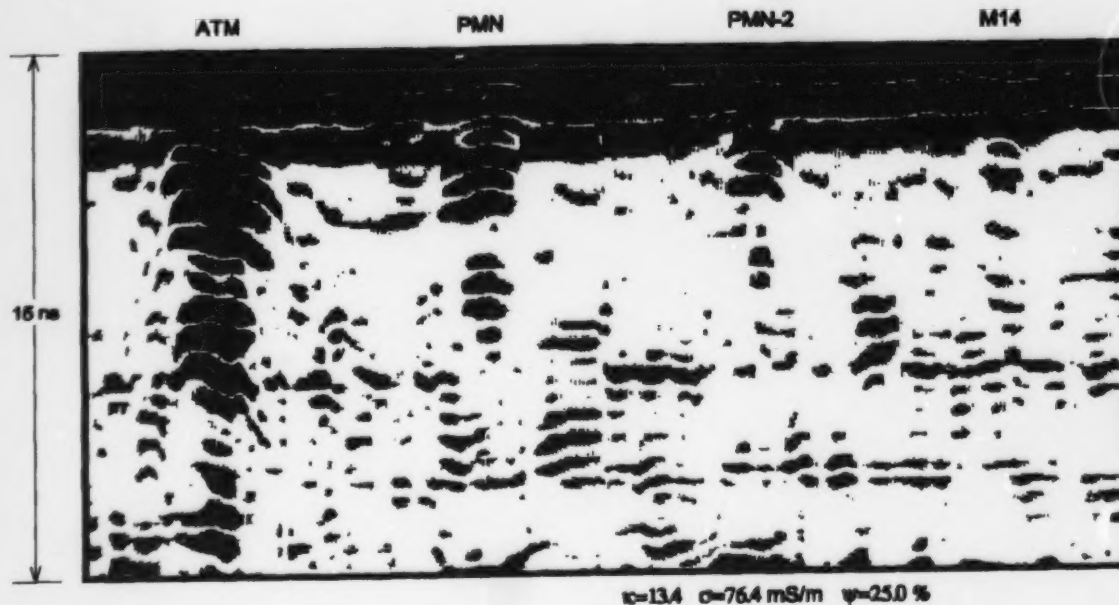
Project No.
 Drawn
 Reviewed



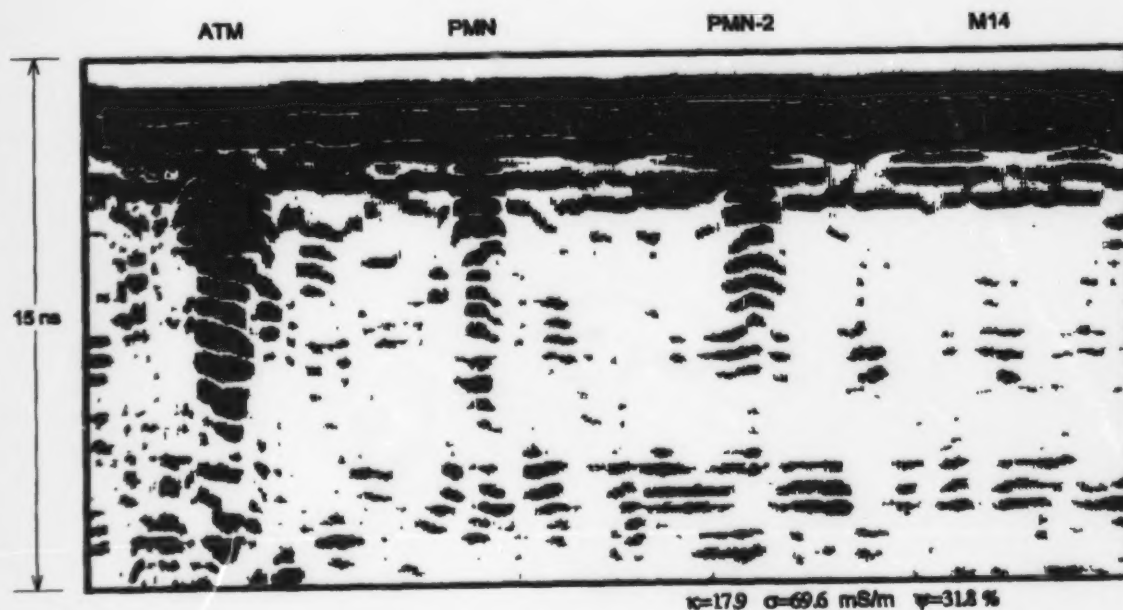
Temperature - Precipitation



VTS-1-1

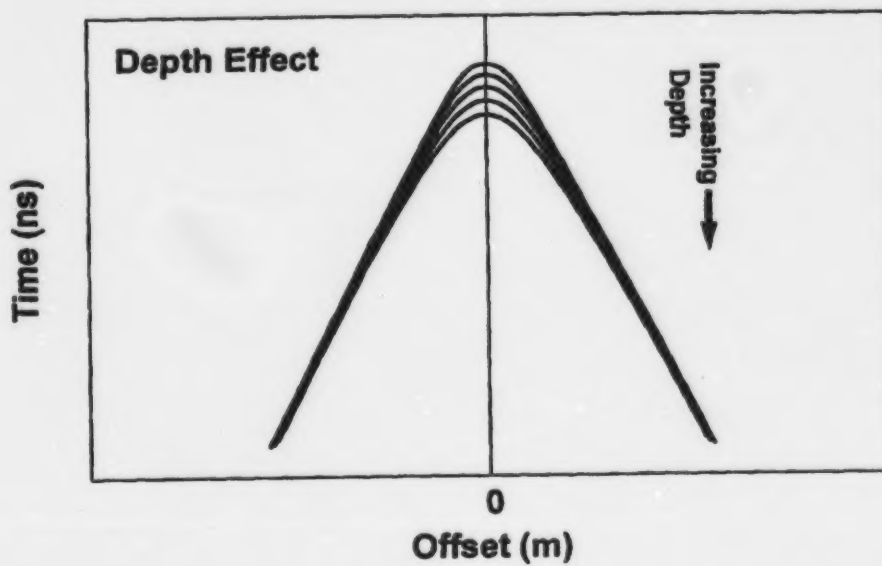
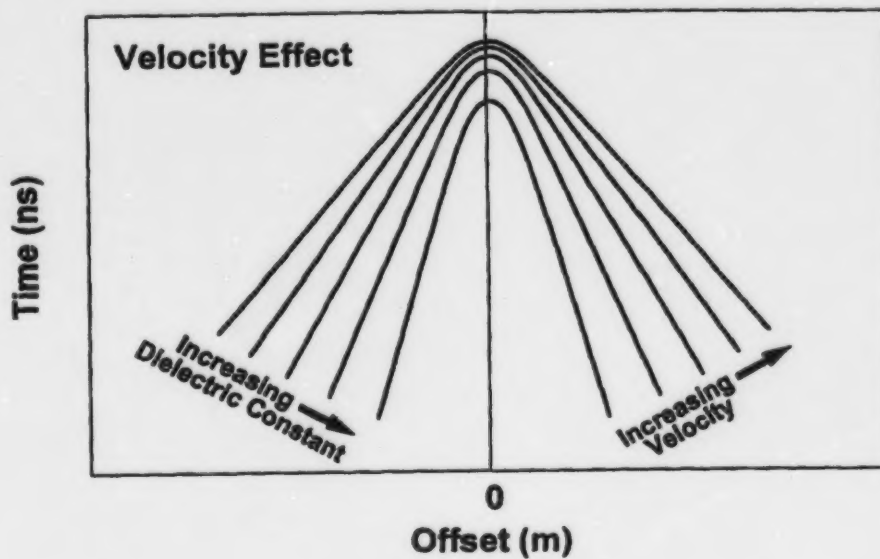


VTS-1-2



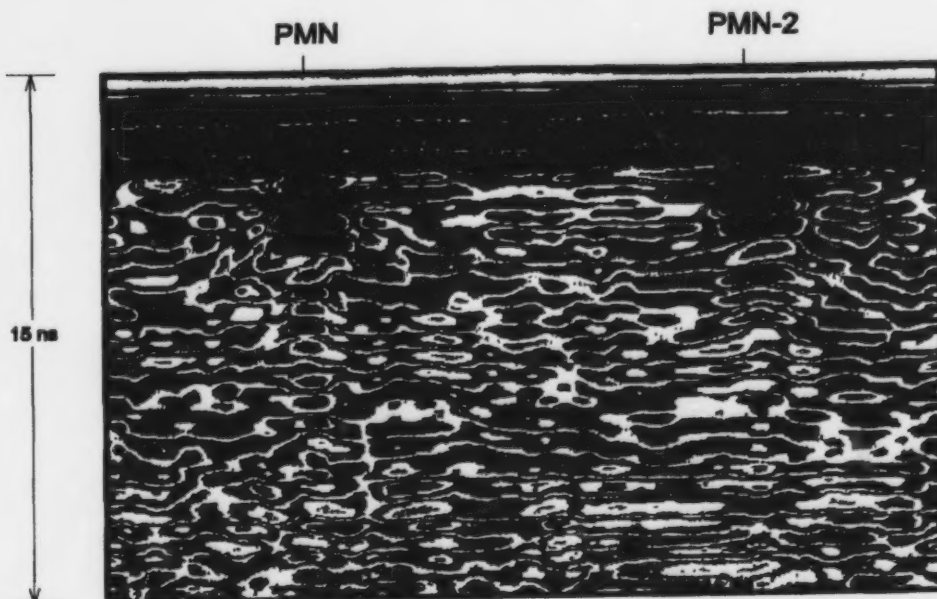
- Centre Frequency: 1 GHz
- Burial Depth: 10 cm

ϵ - Dielectric Constant (TDR 10 cm)
 σ - Electrical Conductivity (15 kHz Geonics EM-38)
 ψ - Volumetric Moisture (Topp et al., 1980)

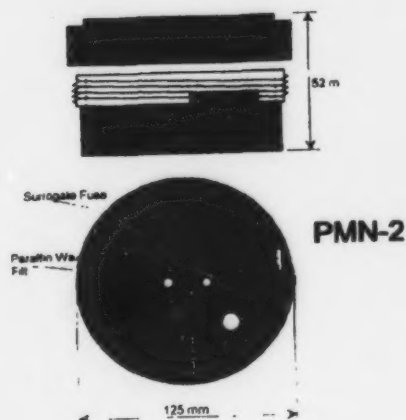
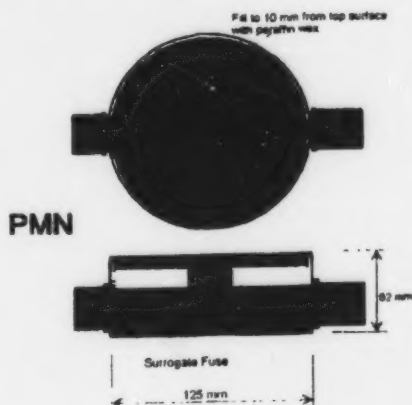


VTS-1-2

- Centre Frequency: 1 GHz
- Burial Depth: 10 cm



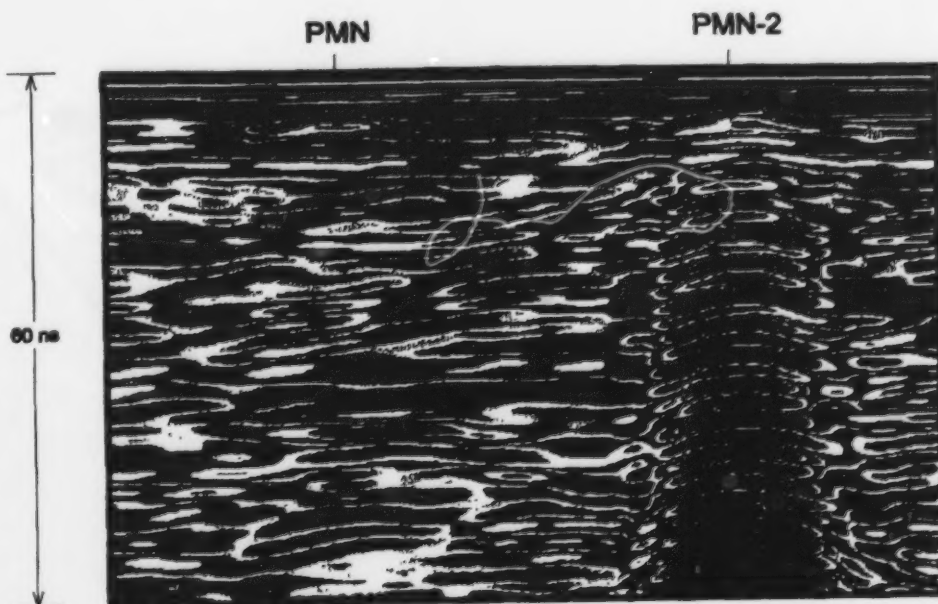
$\kappa=17.9$ $\sigma=69.6$ mS/m $\psi=31.8$ %



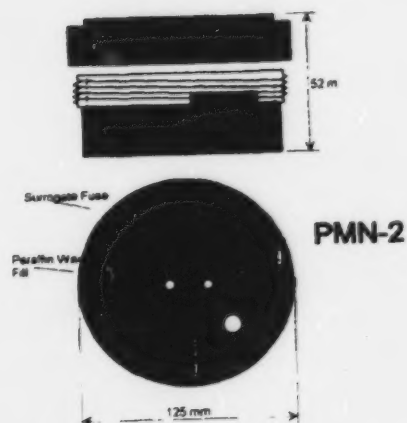
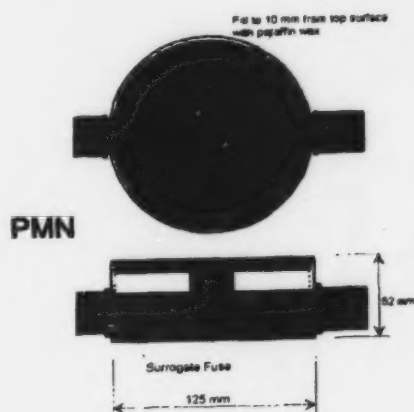
κ - Dielectric Constant (TDR 10 cm)
 σ - Electrical Conductivity (15 kHz Geonics EM-38)
 ψ - Volumetric Moisture (Topp et al., 1980)

VTS-1-2

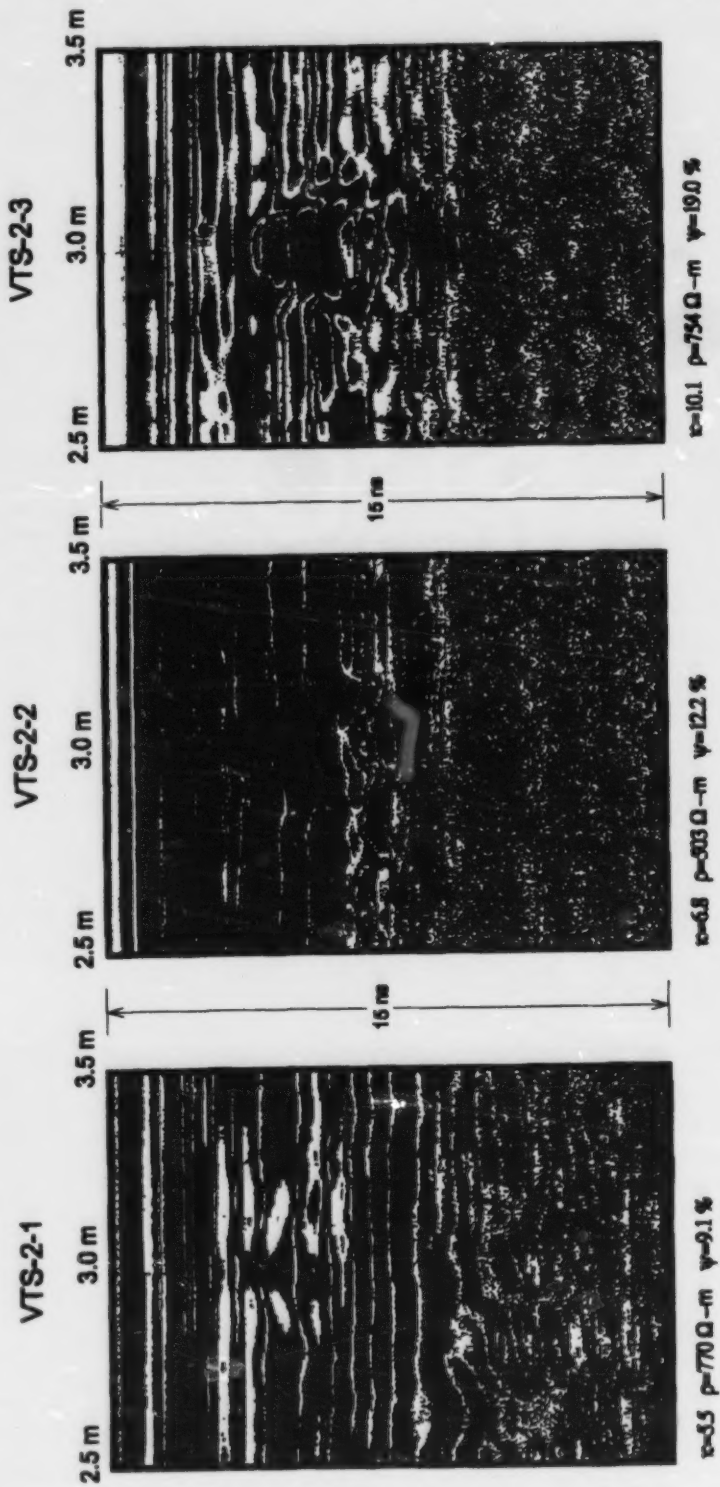
- Centre Frequency: 500 MHz
- Burial Depth: 10 cm



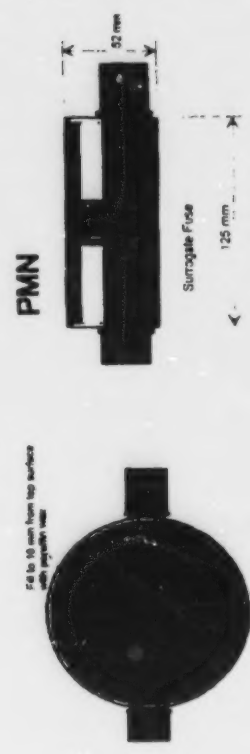
$\kappa=17.9$ $\sigma=69.6 \text{ mS/m}$ $\psi=31.8 \%$

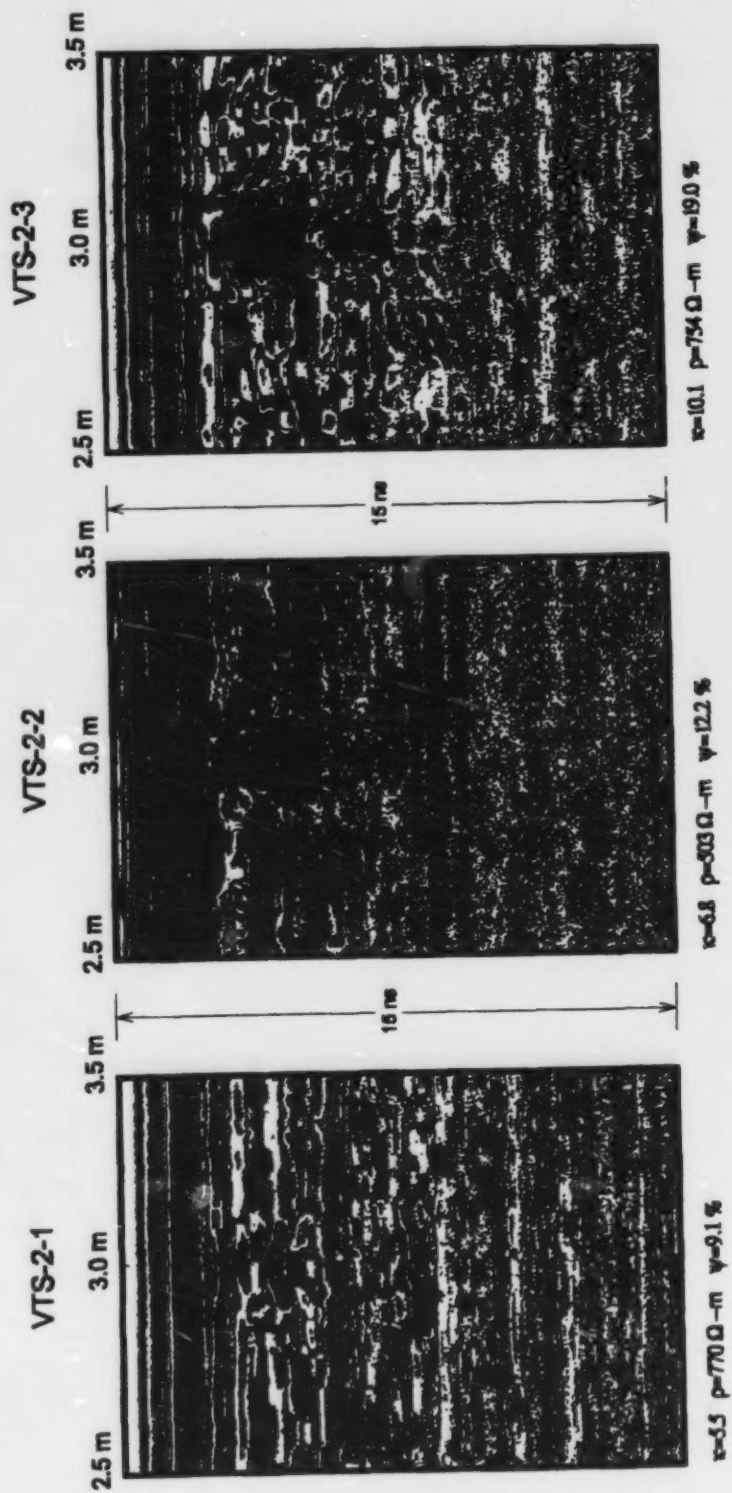


κ - Dielectric Constant (TDR 10 cm)
 σ - Electrical Conductivity (15 kHz Geonics EM-38)
 ψ - Volumetric Moisture (Topp et al., 1980)

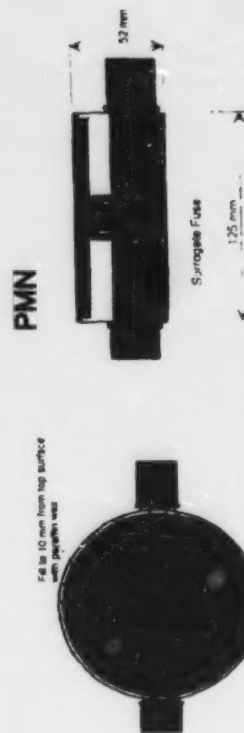


- Surrogate: PMN
 - Centre Frequency: 900 MHz
 - Burial Depth: 10 cm
- κ - Dielectric Constant (TDR 10 cm)
 ρ - Electrical Resistivity (100 Hz Wenner $s=10$ cm)
 ψ - Volumetric Moisture (Topp et al., 1980)



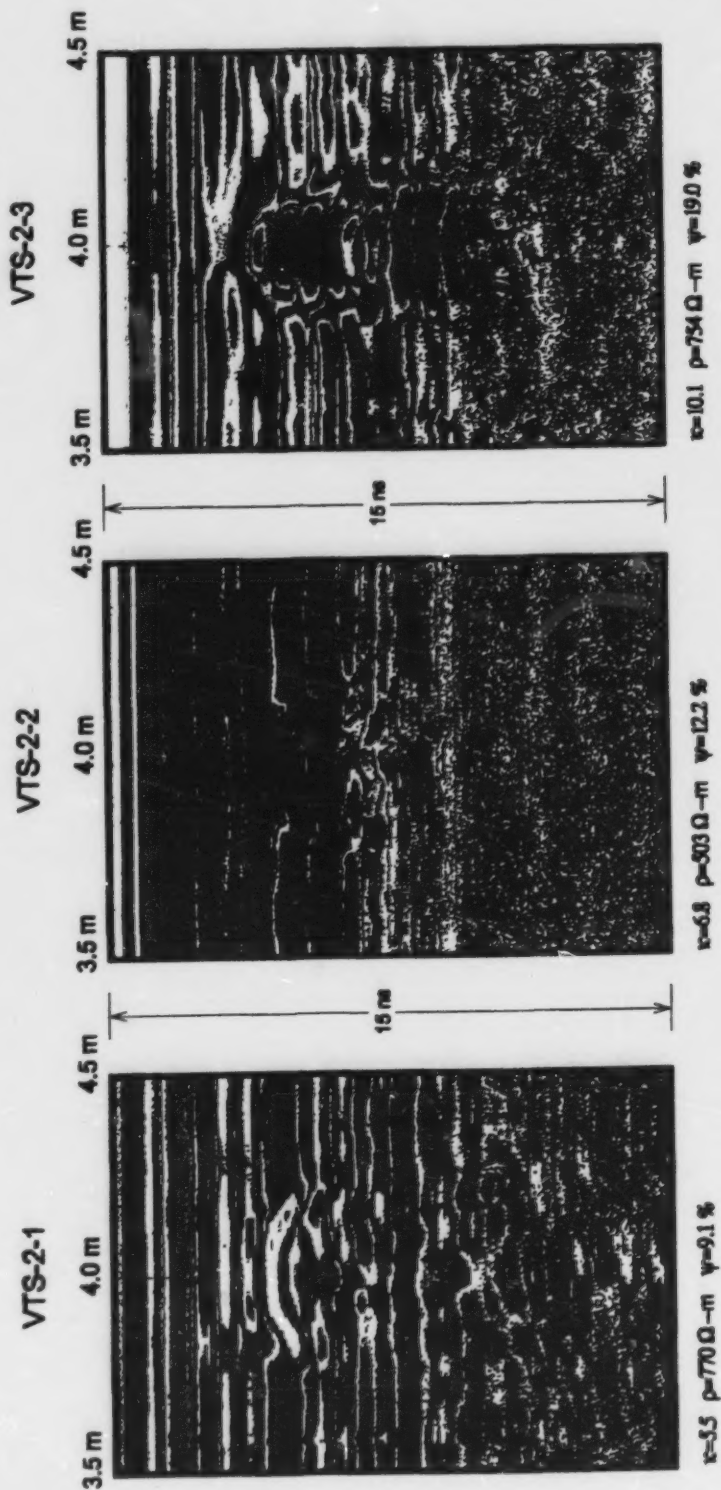


- Surrogate: PMN
 - Centre Frequency: 1 GHz
 - Burial Depth: 10 cm
- κ - Dielectric Constant (TDR 10 cm)
 ρ - Electrical Resistivity (100 Hz Wenner $s=10$ cm)
 ψ - Volumetric Moisture (Topp et al., 1980)

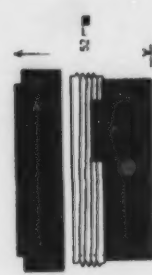


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 Drawn A.G.
 Reviewed G.M.C.

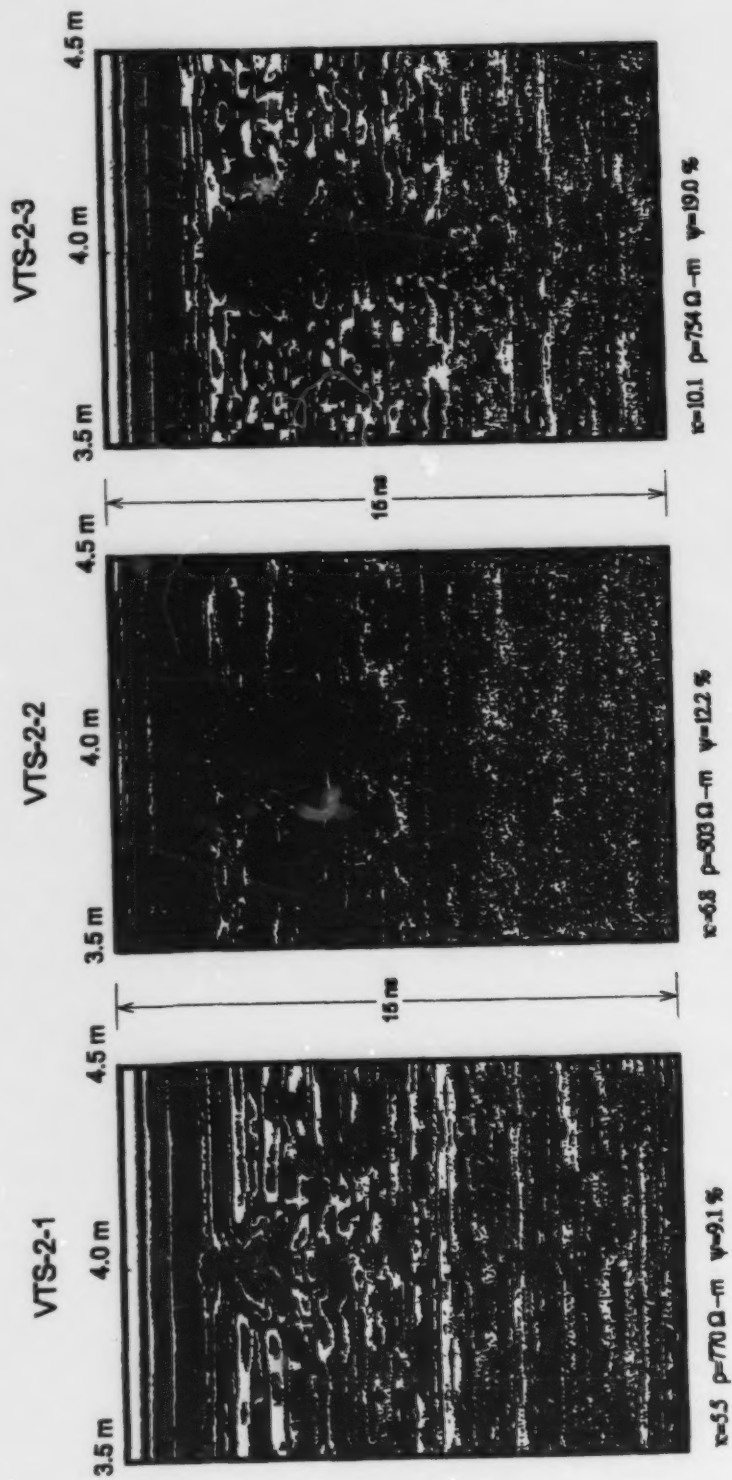
VTS-2 / PMN / 1GHz



PMN-2

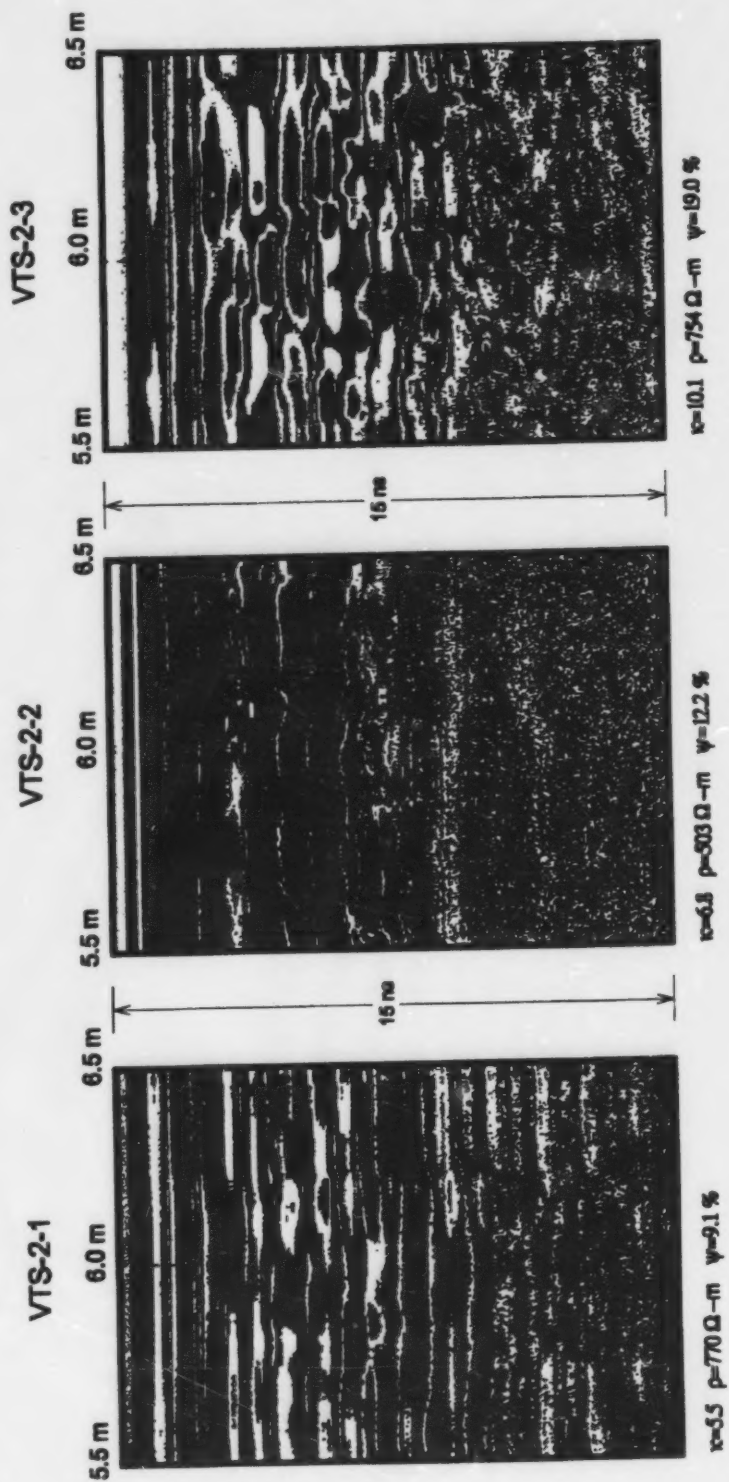


- Surrogate: PMN-2
 - Centre Frequency: 900 MHz
 - Burial Depth: 10 cm
- κ - Dielectric Constant (TDR 10 cm)
 ρ - Electrical Resistivity (100 Hz Wenner $a=10$ cm)
 ψ - Volumetric Moisture (Topp et al., 1980)



- Surrogate: PMN-2
- Centre Frequency: 1 GHz
- Burial Depth: 10 cm

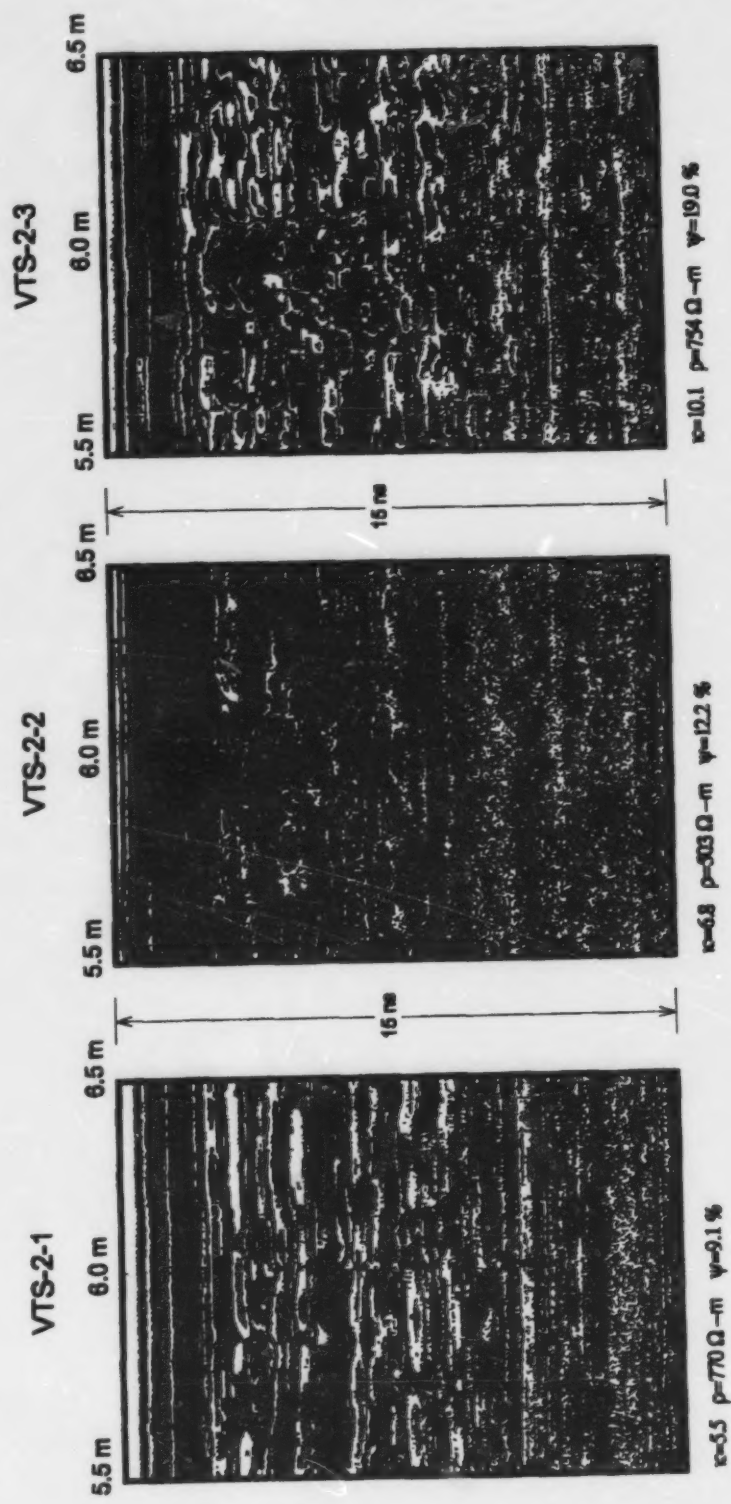
κ - Dielectric Constant (TDR 10 cm)
 ρ - Electrical Resistivity (100 Hz Wenner $a=10$ cm)
 ψ - Volumetric Moisture (Topp et al., 1980)



M14



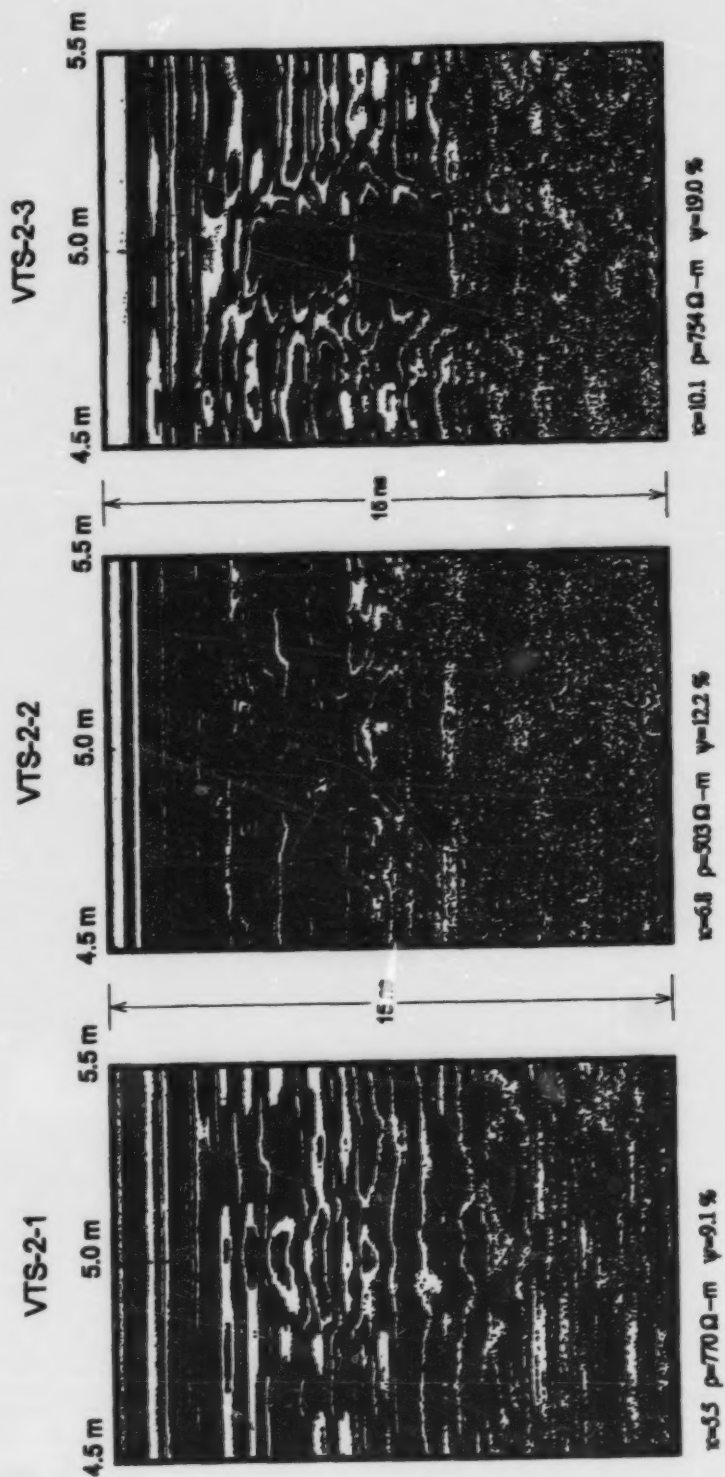
- Surrogata: M14
 - Centre Frequency: 900 MHz
 - Burial Depth: 10 cm
- κ - Dielectric Constant (TDR 10 cm)
 p - Electrical Resistivity (100 Hz Wenner $a=10$ cm)
 ψ - Volumetric Moisture (Topp et al., 1980)



M14

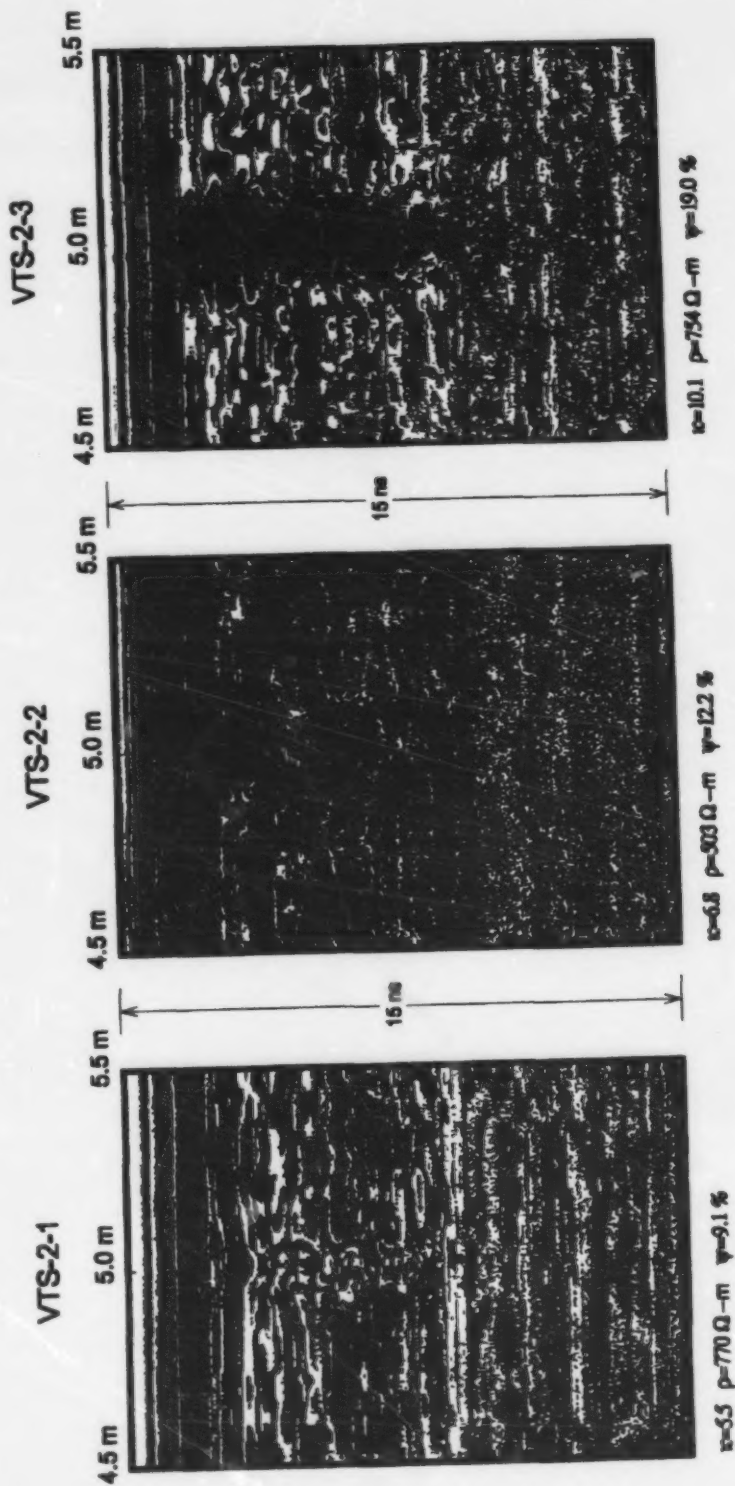


- Surrogate: M14
 - Centre Frequency: 1 GHz
 - Burial Depth: 10 cm
- κ - Dielectric Constant (TDR 10 cm)
 ρ - Electrical Resistivity (100 Hz Wenner $s=10$ cm)
 ψ - Volumetric Moisture (Topp et al., 1980)



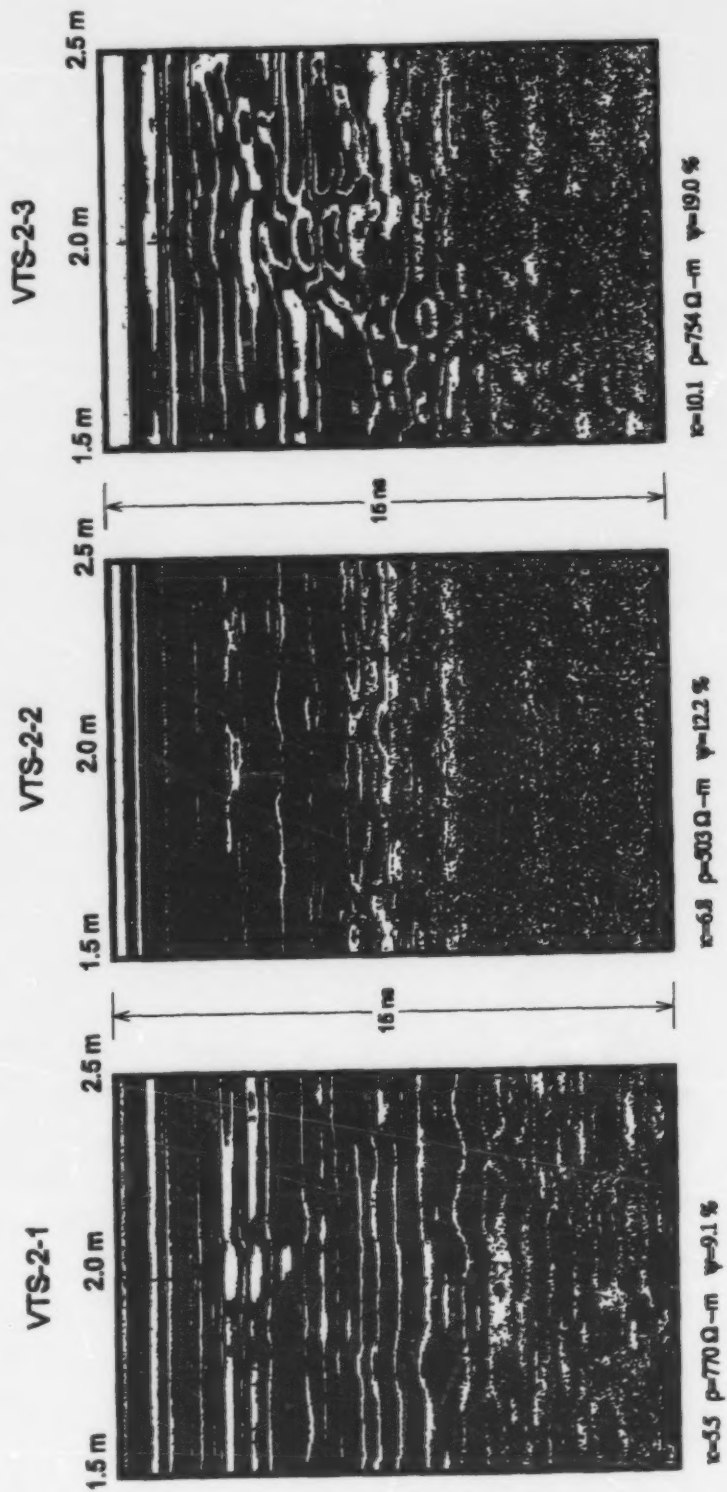
- Surrogate: ATM
- Centre Frequency: 900 MHz
- Burial Depth: 10 cm

κ - Dielectric Constant (TDR 10 cm)
 ρ - Electrical Resistivity (100 Hz Wenner $a=10$ cm)
 ψ - Volumetric Moisture (Topp et al., 1980)



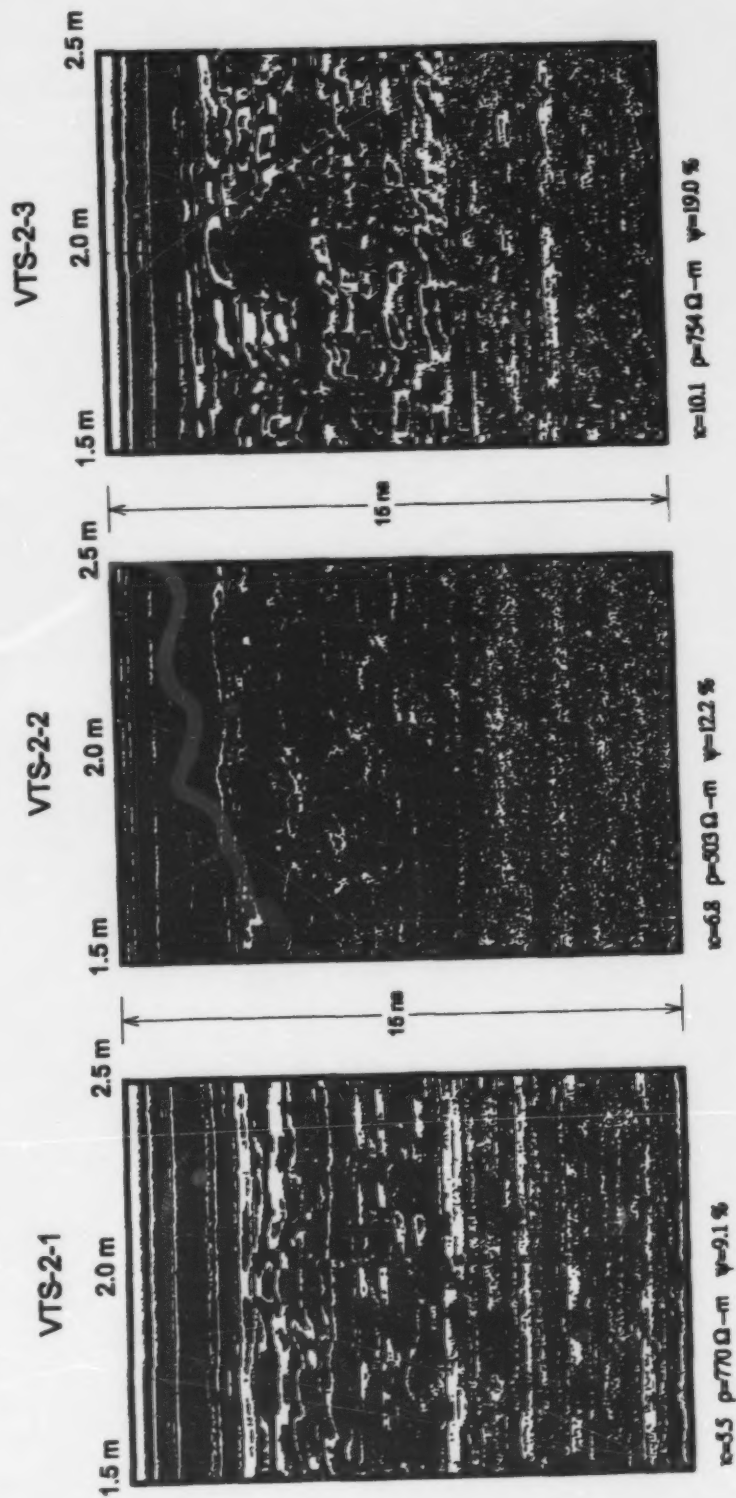
- Surrogates: ATM
- Centre Frequency: 1 GHz
- Burial Depth: 10 cm

ϵ - Dielectric Constant (TDR 10 cm)
 p - Electrical Resistivity (100 Hz Wenner $a=10$ cm)
 ψ - Volumetric Moisture (Topp et al., 1980)



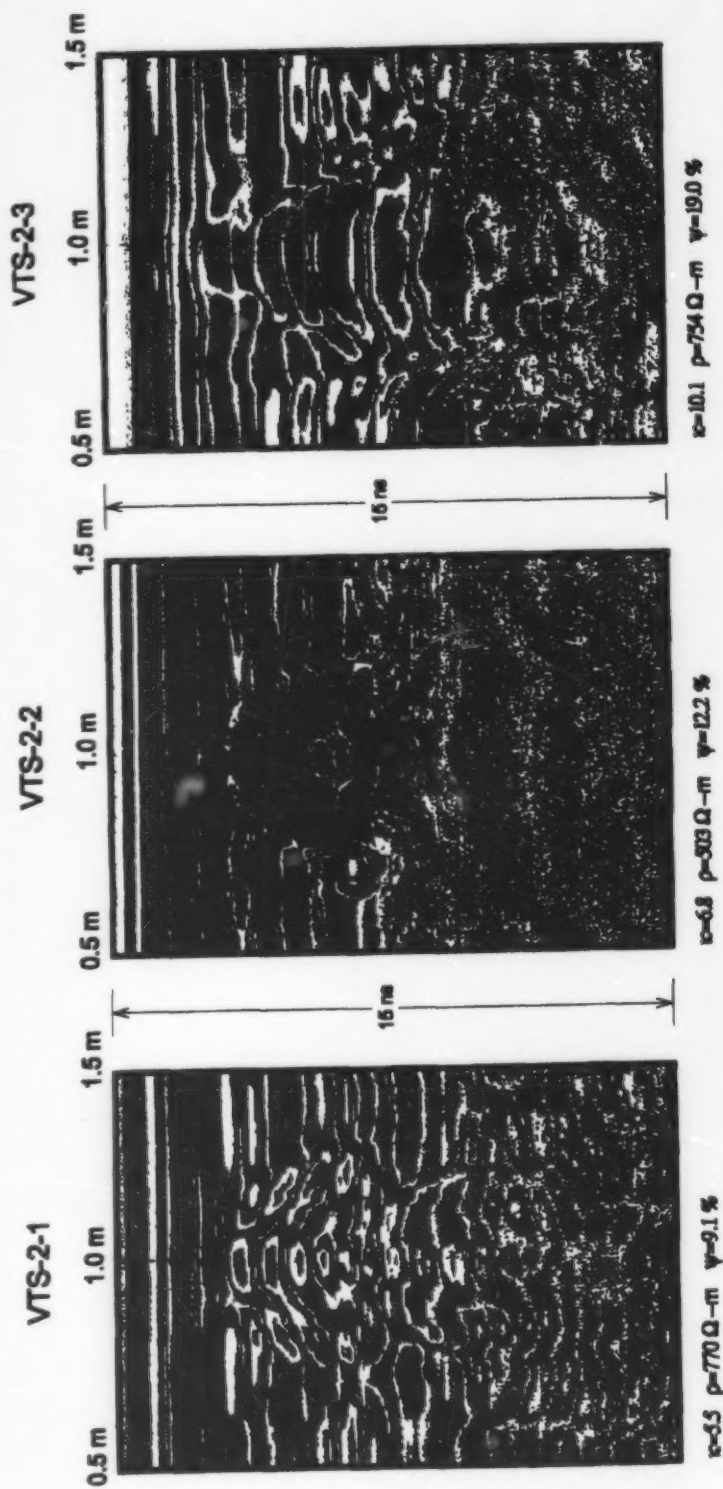
- Surrogate: COBBLE
- Centre Frequency: 900 MHz
- Burial Depth: 10 cm

κ - Dielectric Constant (TDR 10 cm)
 p - Electrical Resistivity (100 Hz Wenner $a=10$ cm)
 ψ - Volumetric Moisture (Topp et al., 1980)



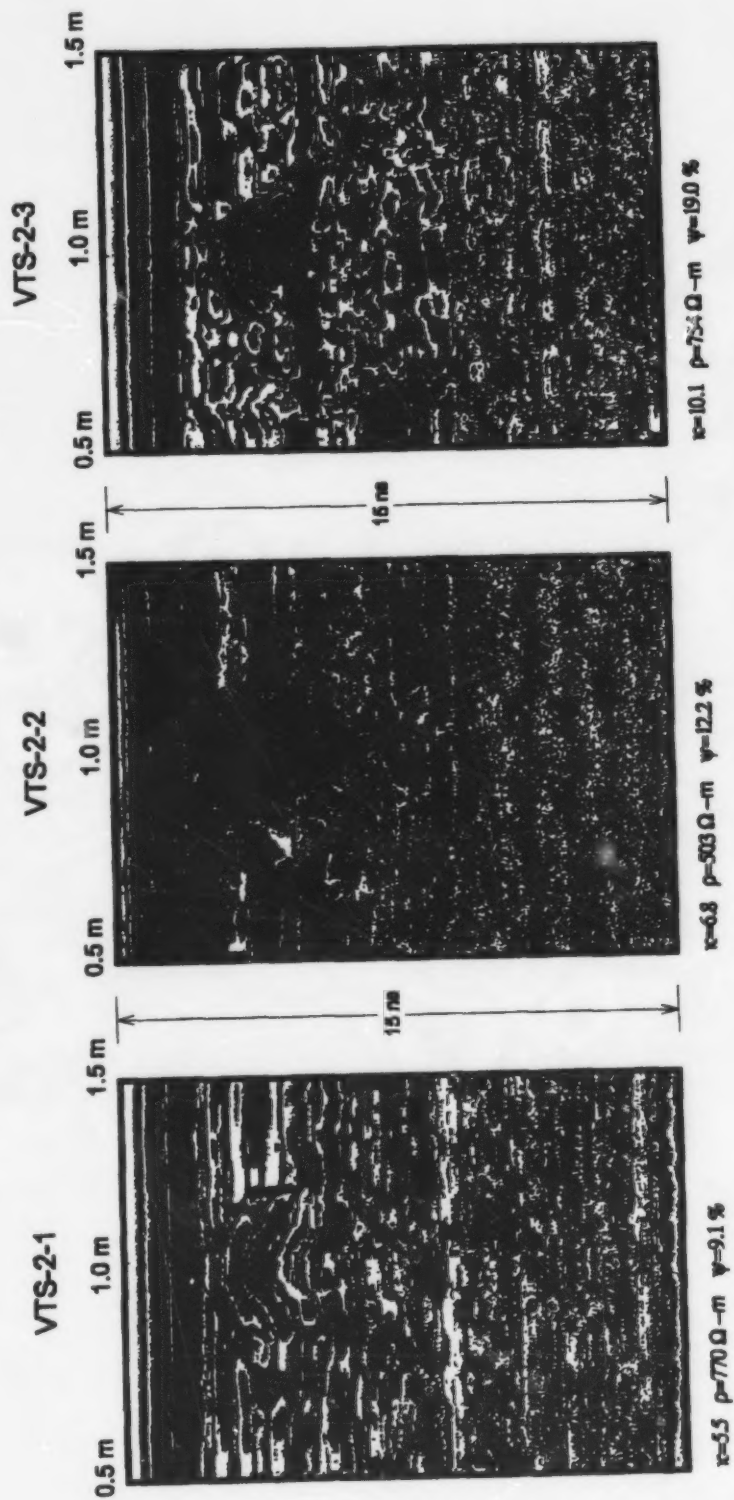
- Surrogate: COBBLE
- Centre Frequency: 1 GHz
- Bural Depth: 10 cm

κ - Dielectric Constant (TDR 10 cm)
 p - Electrical Resistivity (100 Hz Wenner $a=10$ cm)
 ψ - Volumetric Moisture (Topp et al., 1980)



- Surrogate: Ferrous Metal
- Centre Frequency: 900 MHz
- Burial Depth: 10 cm

κ - Dielectric Constant (TDR 10 cm)
 ρ - Electrical Resistivity (100 Hz Wenner $a=10$ cm)
 ψ - Volumetric Moisture (Topp et al., 1980)



- Surrogate: Ferrous Metal
- Centre Frequency: 1 GHz
- Burial Depth: 10 cm

κ - Dielectric Constant (TDR 10 cm)
 ρ - Electrical Resistivity (100 Hz Wenner $a=10$ cm)
 ψ - Volumetric Moisture (Topp et al., 1980)

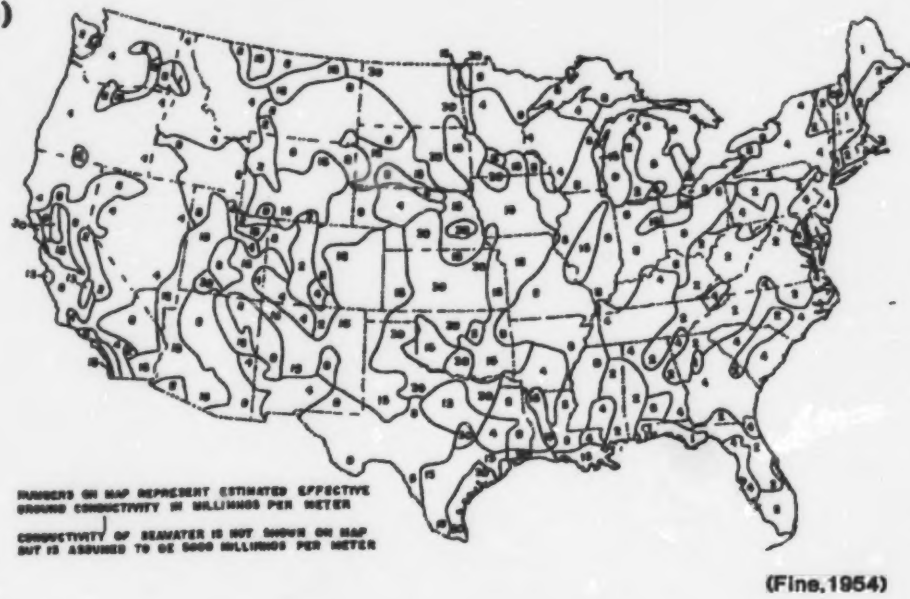
Project No. _____ Drawn _____ Reviewed _____ Date _____

(a)



FIGURE 2-8
Soil Conductivity Map of Canada (Bodie et al., 1976)

(b)



Project No. Drawn Reviewed Date

(a)

POTENTIAL FOR GROUND-PENETRATING RADAR SOILS INTERPRETATIONS

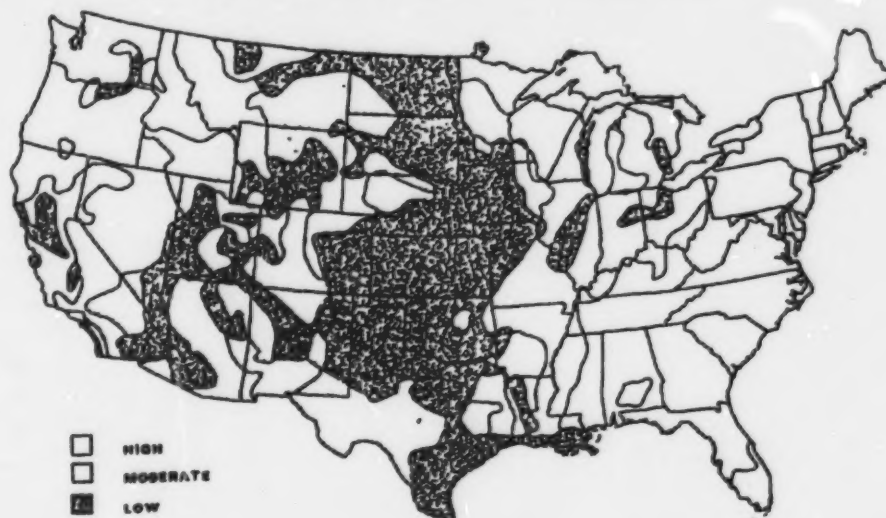


Fig. 2-11. A potential map for GPR applications. (Doolittle, 1987)

(b)

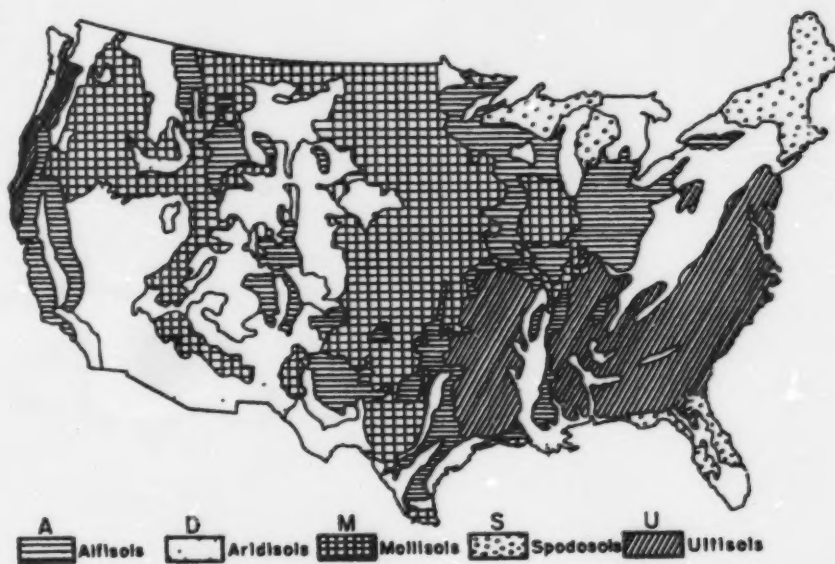


Fig. 19. Generalized map of five dominant soil orders. The blank spaces comprise Inceptisols, followed by Entisols and small areas of Histosols and Vertisols. Oxisols are absent

Project No. Drawn Reviewed Date

(a)

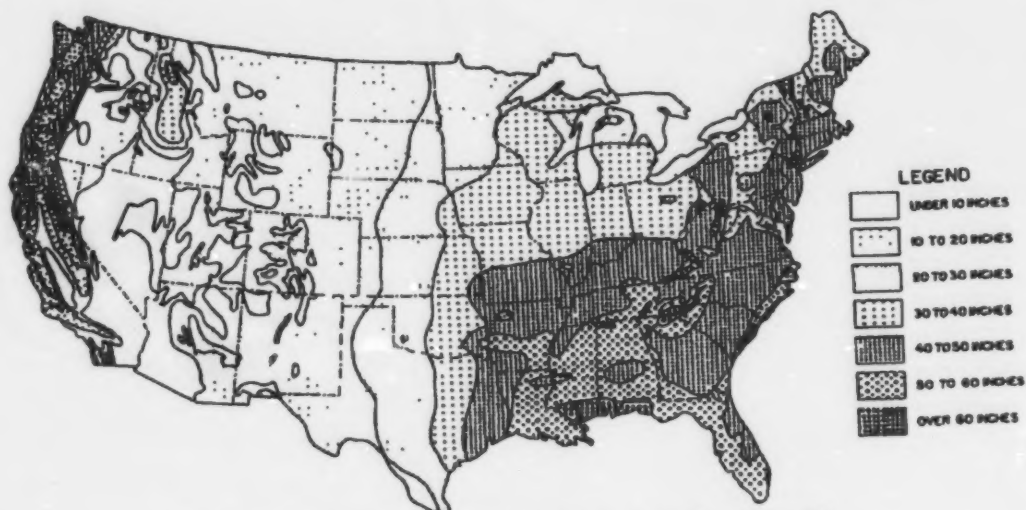


FIG. 50. —Distribution of mean annual precipitation in the United States (44).

(b)



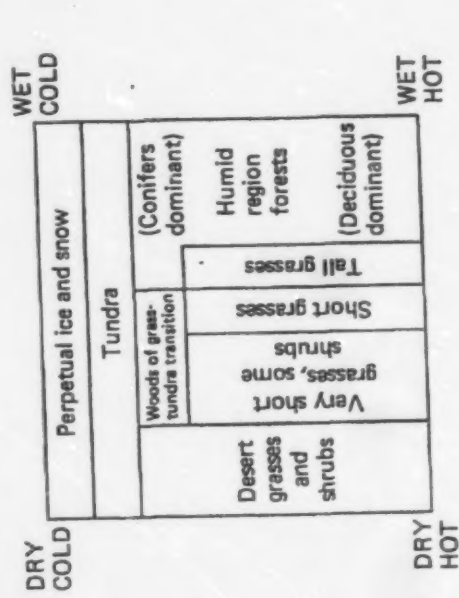
WESTERN FORESTS

- Rocky Mountain
- Pacific Coast
- Grasslands
- Desert types

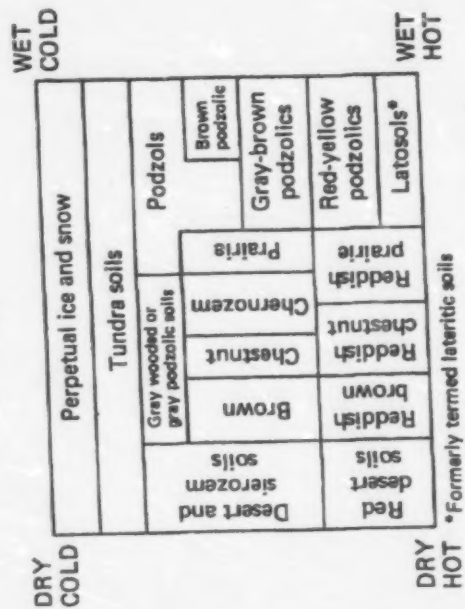
EASTERN FORESTS

- Northern
- Central hardwood
- Southern
- Tropical

(Brady, 1974)



(a)



(b)

(Brady, 1974)

*Formerly termed lateritic soils

Climate-soil relations: Each area describing the climatic conditions under which a given Great Soil Group occurs is delineated by the standard deviations from the means of temperature and biofactor for all samples of that group. (After H. Kohnte, R. G. Shuf, and P. A. Miller, Quantitative Relations between Climate and Soil Formation. "Zeitschrift f. Pflanzenernährung und Bodenkunde," vol. 119, pp. 24-33, 1968.)



(Smith-Rose, 1934)

Site	Geological classification	Depth	Description of sample	Moisture content per cent	Conductivity (in a.s.a.) at 30° C. and at frequencies in kilocycles per sec. of:—					Apparent dielectric constant (in a.s.a.) at frequencies in kilocycles per sec. of:—	
					1	100	1 500	10 000	1 500		
					3.0 × 10 ⁵	3.4 × 10 ⁵	3.9 × 10 ⁵	6.0 × 10 ⁵	100	1 500	10 000
Rugby Radio Station (1)	Lower lias	Surface	Dark fibrous loam	60	3.0 × 10 ⁵	3.4 × 10 ⁵	3.9 × 10 ⁵	6.0 × 10 ⁵	100	100	55
		1	Loam and clay	33	6.5	7.0	7.0	9.0	96	96	43
		2	Clay and sand	26	7.5	8.0	8.0	12.0	105	105	48
		3	Blue clay	25	8.0	9.0	9.5	11.0	95	95	46
Rugby Radio Station (2)	Lower lias	10	Blue clay	23	6.0	6.5	7.0	8.5	80	80	29
		Surface	Loam	22	0.55	0.65	0.70	1.1	35	35	25
		1	Loam and clay	18	0.85	0.85	1.2	1.4	32	32	21
		3	Blue clay	27	6.5	7.5	8.0	13.0	125	125	75
Baldock, Herts (P.O. Receiving Station)	Chalk	5	Clay and sand	21	5.5	5.5	6.5	9.0	75	75	42
		10	Blue clay	25	7.0	8.0	8.5	12.0	70	70	40
		Surface	Fibrous loam	21	0.85	0.90	0.95	1.4	39	39	23
		1	Chalky loam	21	0.55	0.55	0.85	0.95	41	41	25
Tatfield, Kent (B.B.C. Receiving Station)	Upper greensand	2	Chalk	24	0.26	0.26	0.38	0.61	28	28	21
		3	Chalk	27	0.55	0.70	0.80	1.0	35	35	29
		5	Chalk	26	0.91	1.4	1.6	2.0	45	45	38
		10	Chalk	27	1.3	1.4	1.5	1.9	44	44	34
Tatfield, Kent (B.B.C. Receiving Station)	Upper greensand	Surface	Fibrous loam	37	2.7	3.4	4.0	5.0	80	80	49
		1	Brown, sandy clay	19	2.2	2.4	2.4	3.8	39	39	19
		2	Brown sand	15	1.8	2.0	2.1	3.3	33	33	19
		3	Light brown sand	13	1.4	1.5	1.6	2.1	33	33	25
Brookmans Park, Herts (London Regional Station)	London clay	5	Light brown sand	20	2.3	2.3	2.5	4.0	60	60	37
		10	Yellow sand	15	1.5	1.7	2.0	2.9	60	60	26
		Surface	Fibrous loam	19	0.90	0.95	1.1	1.7	48	48	25
		1	Stony loam	18	0.85	0.90	0.95	1.4	40	40	24
Brookmans Park, Herts (London Regional Station)	London clay	2	Light sandy clay	23	0.50	0.65	0.70	1.6	45	45	24
		3	Sandy clay	22	1.1	1.2	1.5	2.9	70	70	33
		5	Sandy clay	21	1.5	1.7	1.9	2.6	75	75	36
		8 to 10	Clay and shingle	10	1.5	1.6	1.6	2.7	43	43	23

Figure

I-1

ELECTRICAL PROPERTIES
CONDUCTIVITY-PERMITTIVITYProject No.
Drawn
Dated.....

Site	Geological classification	Depth	Description of sample	Moisture content per cent	Conductivity (in a.s.u.) at 20° C. and at frequencies in kilocycles per sec. of:—				Apparent dielectric constant (in a.s.u.) at frequencies in kilocycles per sec. of:—	
					1	100	1 300	10 000	1 300	10 000
					0.85×10^8	0.98×10^8	1.1×10^8	1.6×10^8		
Deventry, Northants	Upper lias	Surface	Fibrous loam	28						
		1	Sandy loam	16	0.34	0.34	0.40	0.61	48	30
		2	Brown sand	14	0.29	0.29	0.33	0.46	20	17
		3	Brown sand	6.0	0.10	0.11	0.13	0.19	11	14
		5	Sand and sandstone	8.5	0.075	0.090	0.13	0.22	14	8
Washford Cross, Somerset (West Regional Station)	Red marls	10	Sand and sandstone	24	0.26	0.33	0.44	0.80	47	31
		Surface	Reddish-brown loam	23						
		1	Reddish-brown clay	20	1.5	1.7	1.8	2.3	46	32
		2	Reddish-brown clay	18	1.6	1.7	1.8	2.5	50	33
		3	Reddish-brown clay	21	2.6	2.8	3.1	3.6	80	45
Brendon Hills, Somerset	Devonian	5	Reddish-brown clay	19	4.7	4.9	5.5	8.0	130	85
		10	Reddish-brown clay	15	2.8	3.1	3.3	4.2	55	40
					4.3	4.5	4.5	6.1	70	55
		Surface	Black fibrous loam	210						
		1	Loam and slate	9.0	1.3	1.5	1.8	2.5	90	65
Merrivale, Dartmoor, Devon	Granite	2	Loam and slate	9.0	0.086	0.030	0.040	0.060	12	10
		3	Loam and slate	8.0	0.016	0.016	0.036	0.035	9.0	8.0
		5	Loam and slate	5.5	0.0055	0.0060	0.013	0.015	7.0	6.5
		10	Loam and slate	—	0.0023	0.0039	0.0085	0.016	6.0	5.5
			Slate	—	0.0026	0.0025	0.0092	0.046	9.5	8.0
Merrivale, Dartmoor, Devon	Granite	1	Gritty loam	18	0.12	0.13	0.16	0.18	22	15
		2	Gritty loam	13	0.14	0.14	0.18	0.20	14	13
		3 to 10	Granite	—	0.00090	0.0070	0.028	0.11	12	8.5
		3 to 10	Granite	—	0.00070	0.0050	0.019	0.095	10.0	7.5
		3 to 10	Granite	—	0.00039	0.0021	0.011	0.050	9.5	7.5
Dousland, Dartmoor, Devon	Devonian	3 to 10	Granite	—	0.000015	0.00090	0.0024	0.015	7.0	6.5
		Surface	Loam	47						
		1	Dark brown loam	41	0.55	0.60	0.70	0.85	50	30
		Below 1	Slate	—	0.24	0.24	0.31	0.47	33	27
		Below 1	Granite	—	0.0028	0.0017	0.011	0.050	9.5	8.0
					0.000030	0.00026	0.0016	0.0070	6.0	6.0

Figure

1-2

ELECTRICAL PROPERTIES
CONDUCTIVITY-PERMITTIVITY

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Mooride Edge, Yorks (North Regional Station)	Millstone grit	Surface 1 2 3 5 10	Fibrous loam Dark grey clay Dark grey clay Dark grey clay Dark grey clay Yellow and grey clay	130	1.2 × 10 ⁸	1.6 × 10 ⁸	1.8 × 10 ⁸	3.9 × 10 ⁸	95	40
				60	1.2	1.4	1.6	3.1	75	36
				35	0.70	0.95	1.0	2.4	85	33
				39	1.1	1.5	1.6	3.1	75	36
				19	0.85	1.0	1.2	2.9	70	31
				15	0.31	0.33	0.37	0.75	28	16
Westerglen, Falkirk (Scot- tish Regional Station)	Boulder clay	Surface 1 2 3 5 10	Fibrous loam Fibrous loam Clay and loam Dark grit and clay Dark grit and clay Dark grit and clay	38	0.55	0.65	0.75	1.1	50	20
				30	1.0	1.0	1.1	1.5	60	20
				19	1.1	1.1	1.2	1.7	60	21
				18	0.60	0.70	0.80	1.2	50	19
				18	1.9	2.2	2.5	3.3	75	22
				15	2.1	2.3	2.6	3.2	80	24
Teddington, Middlesex (N.P.L.)	London clay	Surface 1 2 3 5 7 10	Fibrous loam Sandy loam Sandy loam Fine gravel Coarse gravel Fine sand Sand and shingle	26	1.0	1.1	1.2	1.7	40	14
				20	0.65	0.70	0.90	1.1	23	14
				13	0.60	0.65	0.65	1.0	20	12
				6.5	0.50	0.60	0.65	0.85	20	10
				2.9	0.17	0.18	0.21	0.30	12	6.5
				2.6	0.14	0.14	0.15	0.21	7.5	6.0
Wychbold, Droitwich (Midland Regional Station)	Red marls	1 2 3 5 10 20 30 40 50 100 150 200 250 300	Red clay and loam Red clay Red clay Red clay and stones Red clay and stones Red clay suspension Red clay suspension Red clay suspension Red clay suspension Red clay Grey clay and salt Red clay and salt Red clay and salt Red clay and salt Red clay and salt	15	1.5	1.8	2.0	3.3	49	22
				13	3.1	2.9	3.0	4.7	42	23
				14	2.1	2.2	2.4	4.0	60	29
				15	3.9	4.6	4.8	8.5	55	31
				21	3.7	4.0	4.2	7.0	60	30
				31	6.0	7.0	7.5	11.0	65	16
				41	6.0	7.0	7.0	13.0	65	38
				25	5.5	6.5	6.5	10.0	60	13
				27	7.0	7.5	8.5	12.0	60	16
				28	16.0	21.0	22.0	25.0	—	—
				27	—	—	about 100	—	—	—
				24	—	—	about 100	—	—	—
				22	2.3	2.3	2.4	3.4	30	13
				21	7.0	8.0	8.5	17.0	95	24

Figure

I-3

ELECTRICAL PROPERTIES
CONDUCTIVITY-PERMITTIVITY

Project No.
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 Reviewed by

Project No.
 Drawn by
 Reviewed by

DIELECTRIC CONSTANTS OF MATERIALS, FREQUENCY RANGE 50-1000Mhz¹
(Petroy, 1994)

MATERIAL	WATER CONTENT	E _r	#PNT	MEASUREMENT LOCATION	MAX FREQ DEPTH
AIR		1			
METAL		1-2			
COPPER		1			
IRON		1			
PVC		3			
WATER		80-81			
PURE WATER		81			
SEAWATER		81			
ICE		3-4			
ICE		3-4			
FRESHWATER ICE		4			
SEAWATER ICE		4-8			
SNOW FIRN		1.4			
SOIL	DRY	2-6			
SOIL	WET	5-15			
MINERAL SOILS (high in sand/low in clay)	3-20%	6+3	79	Florida, U.S.	500
SANDY SOIL	DRY	2.6 6			
SANDY SOIL		22	cc	Eastern, Ontario, Canada	100
SANDY SOIL	WET	25 23.5			
LOAMY SOIL	DRY	2.5			
CLAYEY SOIL	DRY	2.4			
CLAY SOIL	DRY	3 2.5			
CLAY-LOAM SOIL	10%	10			
LOAMY SOIL	WET	19			
CLAYEY SOIL	WET	15			
CLAY SOIL	SATURATED	15			
LOAM		27.5+5	11	Location Unknown	≈100
V.FINE GRAINED SOIL		36 ?	cc	Eastern, Ontario, Canada	100

ORGANIC SOILS		64+5	3	Belle Glade, Florida, U.S.		500
FROZEN SOIL		3-6				
PERMAFROST		<u>4-8</u>				
		6				
CLAYS		5-10				
CLAY		4-16				
CLAYS		5-10				
CLAY	SATURATED	8-12				
CLAY	SATURATED	8-12				
CLAY		5-6.5	m	Saskatchewan, Canada	10-1000	
CLAY	DRY	2	m	Lab measurements	100	
CLAY	WET	22	1	Lab measurement	500	
RED CLAY	20% SAT	20.5	m	Lab measurement	150-500	
RED CLAY	20% SAT	23+2	m	Lab measurement	50-150	
GLACIO SILT/CLAY	WET	28+3	11	Northern Finland		
CLAY	77% SAT	<u>63</u>	m	Lab measurements	100	
		27				
CLAY	60% methanol	21	m	Lab measurements	100	
CLAY	64% benzene	2	m	Lab measurements	100	
SILTS		5-30				
SILT		9-23				
SILTS		5-30				
SILT	SATURATED	10				
SILT	SATURATED	10				
SILT	WET	16.5+1	2	Fairbanks, Alaska, U.S.	300	
SEDGE TUNDRA SILTY	MOIST	26	1	Bethel, Alaska, USA	500	
SAND 29/CLAY 14	57% SAT	<u>50</u>	m	Lab measurements	100	
		?				
SILT	FROZEN	6.5	2	Fairbanks, Alaska, U.S.	300	
PEAT		50-78				
PEAT		59+8	12	Stormyran, Sweden		
PEAT		58+6	16	Stromsund, Sweden		
PEAT		<u>69+5</u>	11	Dala Mosse, Sweden		
		61.5				

DELTAIC GRAVEL		4.5	m	Brigham City, Utah, U.S.	50	32
GLACIO GRAVEL/SAND DRY		6+2	23	Northern Finland		
ALLUVIAL SAND/GRAVEL		8+1	7	Hanover, NH, U.S.	300	
SAND/GRAVEL	UNSATURAT	4.5	m	Coventry, Conn., U.S.	80	
SAND/GRAVEL	UNSATURT	5	80	Harwich, Mass, U.S.	80	
		5.5				

GLACIAL						
SAND/GRAVEL	SATURATED	10	m	Coventry, Conn., U.S.	80	
SAND & GRAVEL	SATURATED	25	ce	Mirror Lake, NH, USA	80	

SANDS AND GRAVELS PERMAFROST		5-7.5	m	Fairbanks, Alaska	100,500	
GRAVEL	ICY	3-5	3	Fairbanks, Alaska, U.S.	300	
		5				
SAND		4-25				

SAND	DRY	4-6				
SAND	DRY	2.5	m	Lab measurements	100	
SAND	DRY	5	1	Lab measurement	500	
SAND	DRY	3				
SAND	DRY	2-6				
SAND	DRY	4-6				
SAND	DRY	3-5				
QUARTZ SAND	DRY	1.8-6.0	35	Pinellas County, Florida		
SAND	DRY	3-5				
SAND	UNSATURAT	4.5	m	Mansfield, Conn., U.S.	80	
FINE/MED SAND	UNSATURAT	5	m	Coventry, Conn., U.S.	80	
		4				

GLACIO-LACUSTRINE						
SANDS	PARTIAL SAT	9	ce	Lyndonville, Vermont		
SAND 80/CLAY 5	16%SAT	16	m	Lab measurements	100	
SAND 84/CLAY 2	21%SAT	18	m	Lab measurements	100	
SAND	21%SAT	19	m	Lab measurements	100	
		15				

SAND	SATURATED	25				
SAND	SATURATED	30				
SAND	SATURATED	20-30				
SAND	SATURATED	30				
SAND	SATURATED	10	m	Mansfield, Conn., U.S.	80	
SAND	SATURATED	20-30				
SAND	SATURATED	31	m	Hartford, Conn., U.S.	80	
		25				

SAND	18%methanol	10	m	Lab measurements	100	
SAND	19%benzene	3	m	Lab measurements	100	

TILL 9-25

MORaine DRY 9

MAGNETIC PROPERTIES OF TOPSOILS

TABLE 1. Summary of the Test Sites

(Cook and Catts, 1962)

Area, Great Soil Group	No. of Sites	Parent Materials	Range of k , $\mu\text{gauss}/\text{oe}$	Mean Noise, gammas	Textures
N.W. States					
Gray desert	4	Basalt, granite	118-500	10-100	Gravelly
Gray-brown podsol	5	Gabbro, moraines	100-1000	1-100	Organic, silty or cobbles
Brown	4	Sediments or igneous	12-390	30	0 to 30% stones
Gray-brown prairie	2	Basalt, moraine	150-390	6	10%, 2-in. stones
Coast sand	1	Basalt	330	1	Sand (dark)
N. Central States					
Podsol	4	Igneous, moraines	95-600	10-180	30-50% stones
Yellow podsol	4	Sediments	4-97	8	0-20% stones
Prairie	6	Ss, ls, moraines	25-80	1	0-10% stones
Chernozem	2	Sediments, moraine	10-45	1	0-10% stones
Synthetic	2	Sand, iron ore	65-630	...	Fine
S. Central States					
Sands	9	Sediments (?)	0-200	0-1	Uniform, fine
Red desert	10	Rhyolite, limestone	20-356	4	0-30% stones
Redsina	8	Limestone	10-340	6	Gray clays
Reddish-brown	15	Rhyolite, shale	80-520	5-20	Gravelly loams
Reddish-chestnut	17	Sandstone	14-126	2	Fine loams
Reddish prairie	4	Limestone	40-78	3	Crumbly, fine
Brown (etc.)	6	Basalt	83-2200	(10)	70% gravel
Chernozem	3	Loess	61-103	6	Fine loams
Semi-bog	7	Sediments	0-15	2	Clays, silts
Podsolics	5	Sandstone	16-42	2	0 to 80% stones
Solonets	1	Gypsum	15	0	Sandy clay
S.E. States					
Coast sands	11	Sediments (?), coral	0-1	3	Fine beach sands
Red podsol	27	Gneiss, slate	20-2300	(10)	5-20% stones
Brown (?)	72	Granite, sediments	1-1800	...	Generally stony
Panama					
Coast sands	13	Sediments (?), coral	216-150,000	...	Fine (dark or light)
Red humic latosol	7	Basalt, sandstone	200-2000	...	Stony
Reddish-brown humic latosol	6	Tuff, limestone	3-720	...	Magnetised stones
Other latosol	2	Dacite, limestone	425-1000	...	Stony, fine

Magnetic susceptibility of ferri- and antiferromagnetic soil minerals (Mullins, 1977)

Mineral	$10^8 \chi/\text{m}^3 \text{kg}^{-1} (\text{SI})$	Source
Ilmenite (FeTiO_3)	170	1
Hematite ($\alpha\text{Fe}_2\text{O}_3$)	27, 31 up to 63	3 4
Goethite (αFeOOH)	35, up to 126 38, 12.5	4 1, 3
Lepidocrocite (γFeOOH)	50-75 69	4 1
Pyrrhotite (Fe_7S_8) (magnetic pyrites)	5.3×10^3	1
Magnetite (Fe_3O_4)	$5-10 \times 10^4$ $3.9-5.8 \times 10^4$	2 5
Maghemite ($\gamma\text{Fe}_2\text{O}_3$)	$4.1 \times 10^4, 4.4 \times 10^4$	2

